THE APPLICATIONS OF REMOTE SENSING TO GEOMORPHOLOGICAL NEOTECTONIC MAPPING IN NORTH CANTERBURY NEW ZEALAND

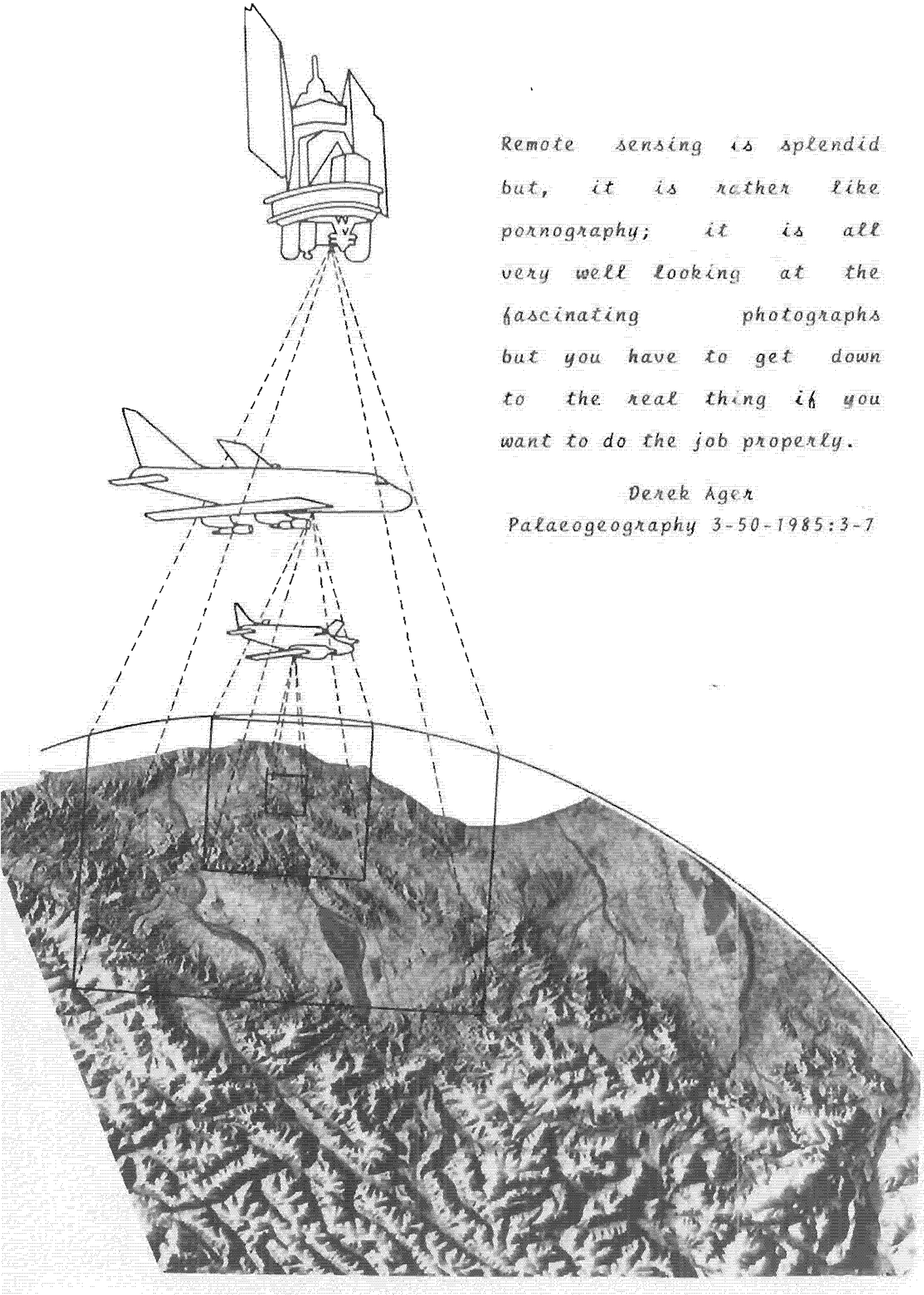
A thesis
submitted in partial fulfilment
of the requirements for the Degree
of
Doctor of Philosophy in Geology
in the
University of Canterbury
by
Hekmat Subhi Yousif

University of Canterbury
1987

بسم الله الرحمن الرحيم

In the name of God, the Merciful, the Compassionate

DEDICATED
TO MY
MOTHER AND FATHER
WHOSE SACRIFICE HAS MADE
THIS POSSIBLE.



Remote sensing is splendid but, it is rather like pornography; it is all very well looking at the fascinating photographs but you have to get down to the real thing if you want to do the job properly.

Derek Ager
Palaeogeography 3-50-1985:3-7

ABSTRACT

Techniques of remote sensing and air photo interpretation have been applied in an investigation of the influence of neotectonism on geomorphic development within the Waipara Region, North Canterbury, New Zealand.

The present physiography is viewed as the product of a series of interactions between fluvial and marine processes operating on the underlying geology that has been subjected to both past and ongoing structural deformation by folding and faulting. The principal landforms can be divided into smaller units controlled by rock induration, modes and rates of weathering, and changing climatic and tectonic regimes. Lithology and bedrock structure influence geomorphology and drainage patterns by controlling relative erodibility of rocks and stability of slopes. Tectonic aspects of the late Quaternary geomorphology of the Waipara region have been studied by mapping the distribution of the late Quaternary surfaces formed by both marine and fluvial processes.

Sequences of marine terraces, that have at most five distinct levels, are well preserved along the eastern margin of the study area. The height and aerial distribution of former shoreline features vary markedly. Both rates and amounts of uplift of individual terraces are variable. Consequently, the present heights of the marine terraces are regarded as the superimposed combination of eustatic sea level change and tectonic movement.

The origin of the lower Waipara gorge is interpreted in terms of an hypothesis of antecedence, linked to at least seven periods of tectonic and climatic base level change. Significant recognizable geomorphic features include: fluvial aggradation (Teviotdale and Canterbury Surfaces), last Glacial and Interglacial marine terraces, lateral channel migration (ancestral and present Waipara River), and the modern flood and coastal plains. The same applies to the smaller adjacent Carrington, Yellow Rose and Kate stream systems and the Teviotdale River.

Terrace lines, resulting from changes in mode of stream operation, provide isochronous features that can be applied to tectonic analysis of the study region. Surveying of tilted straths and warped aggradation terraces, together with other geomorphic evidence, indicate not only active growth of the main Cass Anticline during the late Quaternary, but also that the sites of most active deformation have migrated more recently to the younger flanking structures of the Kate and Black Anticlines.

New Zealand rivers provide good examples of the nature of channel response to actively growing folds and faults and despite differences in size and erosive power, they all show features in common. The evolution of their fluvial systems is detailed in the geomorphic maps prepared from aerial photographs and Landsat imagery at various scales. The main features of channel response to uplift are summarized below:

1. Profiles of tilted straths show conclusive reversals of tilt on the upstream side and an oversteepened gradient on the downstream side of the uplift.

2. The upstream side of the uplift has a higher sinuosity index than on the side immediately downstream.

3. The meandering valley topography has been accentuated by incision of the channel pattern leaving relict straths surfaces preserved at higher levels.

4. A significant rotation in the general alignment of the valley axes relative to their present positions has taken place, typically switching abruptly from synclinal troughs across adjacent anticlines.

5. In uplifted areas that have been affected by accelerated uplift and tilting, cross-valley morphology consists of two distinct parts. An inner (topographically lower) steep sided, V-shaped profile is incised on the older, much more broadly U-shaped valley morphology.

6. Aggradation has occurred in the reaches upstream and downstream of the uplift.

7. The evolution of slopes and drainage and the hydrologic character of individual rivers appear more sensitive to the relatively slow processes of tectonic uplift and tilt than has been generally realized.

8. Irregularities in the shape of longitudinal river profiles reflect major active geologic structures, indicating recurring uplift.

These observations have been used to develop a model for the evolution of drainage within a belt of folding induced by deep seated shear. This model is used to develop general principles for the production of neotectonic maps in other areas of only strike slip tectonics.

CONTENTS

VOLUME ONE

CHAPTER FOUR: DESCRIPTIVE GEOMORPHOLOGY OF THE STUDY AREA

CHAPTER FIVE: MORPHOMETRIC ANALYSIS

LIST OF FIGURES

LIST OF TABLES

LIST OF STEREOMODELS

NOTE:
All stereomodels are at scale of 1:25,000. For over view refer to Plates 1 and 2. A pocket stereoscope is required to obtain the three-dimensional image.

CONTENTS OF MAP ENCLOSURE

VOLUME TWO

PLATES

CHAPTER ONE

1.1 INTRODUCTION

During the late Quaternary, the interaction of tectonism and climate has had a profound effect on the development of the New Zealand landscape. This thesis seeks to explore the impact of these two factors in relation to the evolution of the geomorphology of the Waipara region.

The Waipara area (Fig. 1.1) is located just to the south of the Marlborough Shear Zone which has its origins in the early Middle Miocene as a belt of active folding and faulting exposing Torlesse Group basement rocks and Cenozoic covering strata (Carter & Carter 1982).

Throughout the study area a series of en echelon folds is associated with the southwest-northeast dextral shear regime, across a broad zone extending from the coast to the foothills of the Southern Alps. Here folding commenced in the Late Pliocene which suggests that deformation of the Marlborough Shear Zone spread progressively southwards with time. In response to increasing angular shear, these folds rotated in sigmoidal fashion and were accompanied by a developing pattern of faults which broke through the overlying cover as shear strain accumulates. The width of the folded belt also increased progressively with time.

The Waipara River cuts across the area and, together with local streams, is deeply incised across the fold and fault traces. The associated river terraces are key elements in understanding the local and regional history. The sequence of terraces formed by the Waipara River and Carrington Creek, together with other elements of the

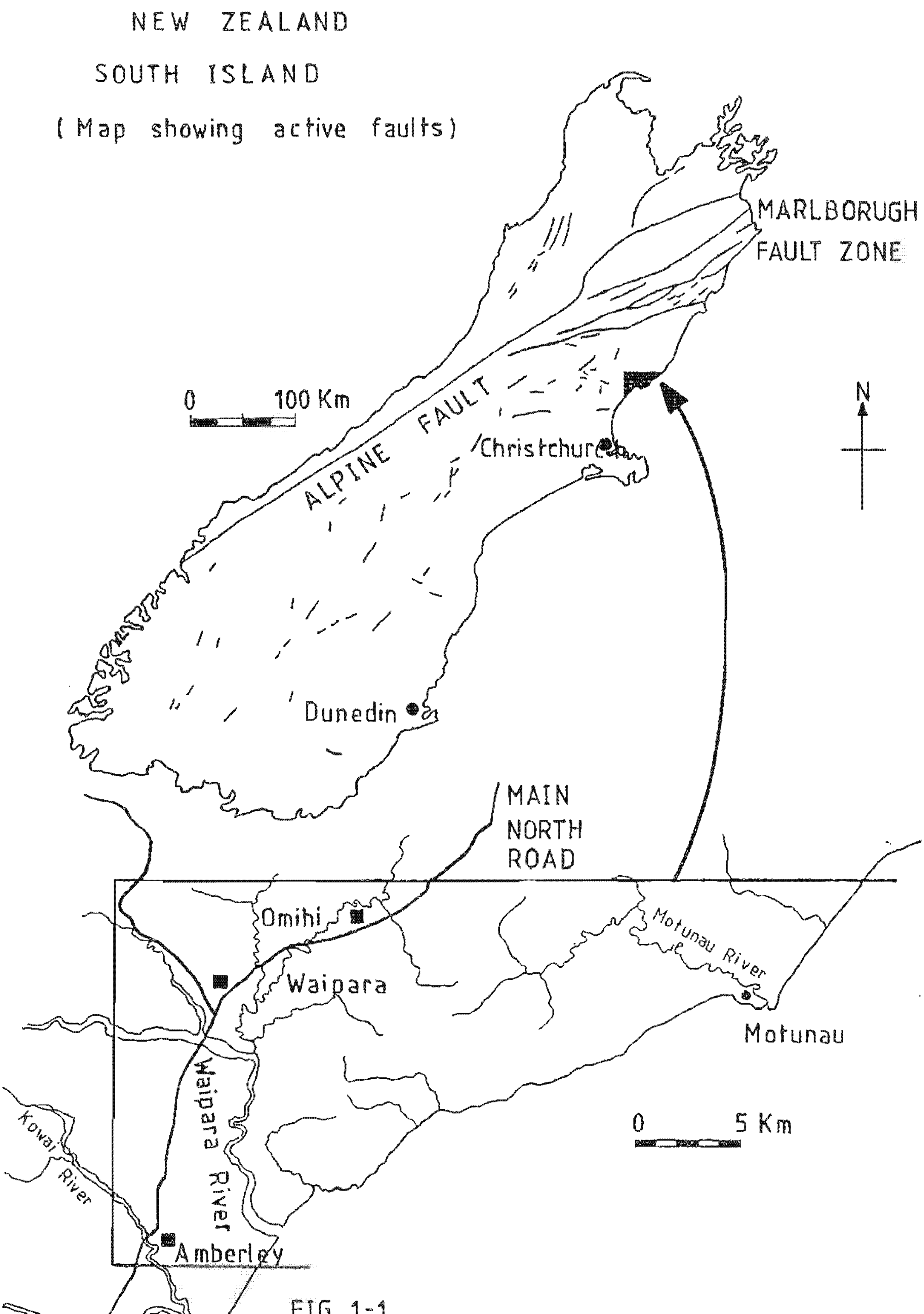


FIG 1-1
The location of the study area

landscape, were mapped using aerial photographs and ground surveyed height control on a scale of 1:10,000 and 1:5,000 respectively.

1.1.1 PURPOSE AND SCOPE OF STUDY

New Zealand is located across one of the major crustal plate boundaries of the world (Pacific-Australian plate boundary zone) and is, predictably, a tectonically active country. Manifestations of the tectonism are earthquakes, volcanic activity, uplifted marine terraces and active faults and folds (Bibby, 1981, Smith 1983, Cole & Lewis 1981, Berryman 1979 & 1980, Hull & Berryman 1986, Suggate, 1978, Wellman 1967 & 1971, Ghani 1978, Lensen 1968, Pillans 1983, Ota et al. 1984, Yoshikawa et al. 1980, Bull & Cooper 1986).

Classical examples of ground deformation surveys are found in New Zealand and California, areas where both the deformation rates are high and the deformation itself is restricted to a relatively narrow zone along major faults (NGS 1973, Walcott & Bibby 1984). Ground deformation studies, employing geodetic triangulation and levelling, provide useful data in the form of changes of angle or distance between stations but tell little of how that deformation is occurring and where it is concentrated. To identify and classify more precisely such relative deformational activity, there is need for a complimentary system of geomorphic mapping. The development of landforms in tectonically active areas is the subject of considerable research but with most of the emphasis, to date on the effects of active faulting. The evolution of drainage in association with active folding is much less well understood with few published papers on the topic. The examples discussed in the literature of tectonic control of river behavior deal primarily with rather dramatic movements, earthquakes, and with the influence of rocks of different resistance emplaced along the channel (Schumm et al. 1982). With recent technological advances and ever increasing coverage air photo and remote sensing techniques are ideally suited for addressing this problem particularly in relation to primary mapping and the targeting of specific locations for more detailed, ground-based field observation.

Some of the most significant advances in geomorphological theory are not yet reflected in morphotectonic studies (Doornkamp 1986). For example, whilst river terraces are accepted by most workers in the neotectonic-morphotectonic field as an indication that base level has fallen or that river incision has been a response to tectonic uplift or tilting, it has been shown that river terraces can be created by changes in sediment load without any external influence such as tilting or a fall in base level (Schumm 1979, Lewis 1944). Terrace formation or, more precisely, cut and fill processes (Leopold & Miller 1957), may also occur in response to human intervention (Cooke & Reeves 1976), or to variations in rainfall, especially in semi-arid lands (Cooke 1974).

The area described in the present thesis lies mainly in the eastern Waipara region (Fig. 1.1) and is known to be one of active folding and faulting. It was selected because the geology, geomorphology and soil of the area are complex, resulting in many different terrain types. As such it provides a useful test of the application of Landsat imagery and aerial photographic interpretation at various scales.

The area lies within a section of the Christchurch-Karamea geodetic traverse from which future data on modern deformation will be forthcoming. Pleistocene marine terraces bordering the eastern margin of the area cut across late Tertiary and Quaternary active geologic structures. These are expected to provide stratigraphic datum points for developing the chronology of drainage evolution. A considerable body of stratigraphic data already exists on the age of aggradation surfaces and Plio-Pleistocene gravel units on the plains along the south and west margins of the area, and these are also expected to be of value in developing a chronology.

Fundamental underlying geological features such as bedding and lithological composition show up well in the geomorphic land form and enable ready recognition of the dominant structural characteristics. The stratigraphic succession is exposed on both limbs of the anticlines offering a wide variety of lithologies and recurrence of these lithologies in differing attitudes and aspects, so that it becomes possible to evaluate the influence and extent of bedrock control on geomorphic processes.

1.1.2 AIMS OF RESEARCH

This thesis describes the application of aerial photographs and Landsat imagery techniques to the interpretation of neotectonic geological problems. In addition, digitizer analysis, ground surveyed height data and photo interpretation procedures were used to construct geological and geomorphological maps of the study area. In particular geomorphic features are used to document the tectonic history of actively growing folds. The ultimate goal was to develop a comprehensive cartographic system applicable to a variety of tectonic environments. Special attention was directed toward:

1. Delineation of the major structural elements.
2. The mode of active deformation.
3. Relative rates of deformation at different places.
4. Absolute rates of deformation where applicable.
5. Late Quaternary stratigraphy.
6. Detailed study of individual geomorphic units and landforms from dynamic, tectonic and morphogenetic aspects.

1.1.3 ORGANIZATION OF THESIS

This thesis is divided into eight chapters. The first and last chapters are introductory and summary chapters, respectively. The second chapter deals with the stratigraphy. This establishes a basis for separating characteristics of a lithological nature from tectonic factors. Chapter Three and Four describe, respectively, the structural and geomorphic characteristics of the coastal range of the Waipara region. This area amply demonstrates the effects of rock structure on landforms. In Chapter Four a review of neotectonic literature, categories of geomorphic data pertinent to tectonic interpretation, and geomorphic classification and cartographic schemes are also discussed. Chapter Five involves several geomorphic concepts and quantitative techniques useful for analyzing drainage patterns and slope development as a possible means of determining relative positions of sea level stands using gradient index parameters. It also reports on the criteria for separating the influence of lithology from tectonic/eustatic controls on base level. Chapter Six describes several geomorphic criteria and techniques useful for identifying such unstable areas and for estimating actual and relative rates of ground deformation; it discusses geologic-geomorphic data that can assist in the formulation of a general model of the evolution of drainage and landscape features associated with active tectonic folding and faulting. This chapter involves seven case studies chosen for contrast in the dominant modes of deformation, catchment size, and erosive power, where geology is already reasonably well documented and in two of these areas a detailed ground base levelling survey was made. Chapter Seven outlines those diagnostic criteria applicable to remote sensing methods tested on areas which have not been subject to previous study and to develop available methodology for the rapid production of small-scale geological and geomorphological maps from Landsat imagery. Five appendices provide data tables, profiles and descriptions of analytical methods employed in this study, plus the writer's publications as conference papers and as published field guides. Microfiche copies of the above data analysis are given in Appendix 1,2,3,4 & 5 (back pocket) of this thesis.

Some of the thesis work has already been formally published. Sections in Chapter Six are papers either submitted for publication (Yousif, in press) or previously published (Campbell & Yousif 1985). Previous works of the late Quaternary geology that might normally be discussed in the introduction are dealt with in individual sections, but repetition is, however, unavoidable as specific areas dealt with in separate publications required a similar introduction to the field area and its geological setting.

1.1.4 LOCATION OF STUDY AREA

This thesis is concerned mainly with the north eastern half of the the Amberley and Motunau sheet (S68 & S69) of the NZ.M.S. 1 topographical map series (1:63,360) in the Waipara region, some 70km due north of Christchurch (Fig. 1.1). It examines an area of about (225 km^2) and extends from Amberley beach in the south to the north of Motunau River. The western and eastern boundaries are the Canterbury Plain and the sea respectively.

1.1.5 FIELD WORK

Initial mapping of the area was spread over a period of two months i.e. February and April 1983. Aerial photographs (1:25,000) were used in the field as base maps on which both lithological and geomorphological features were drawn. In the main, the study area is extensively farmed, with the only significantly populated areas being the Waipara and Amberley townships along the main North Highway Road and small isolated farm settlements situated along the coastal hills.

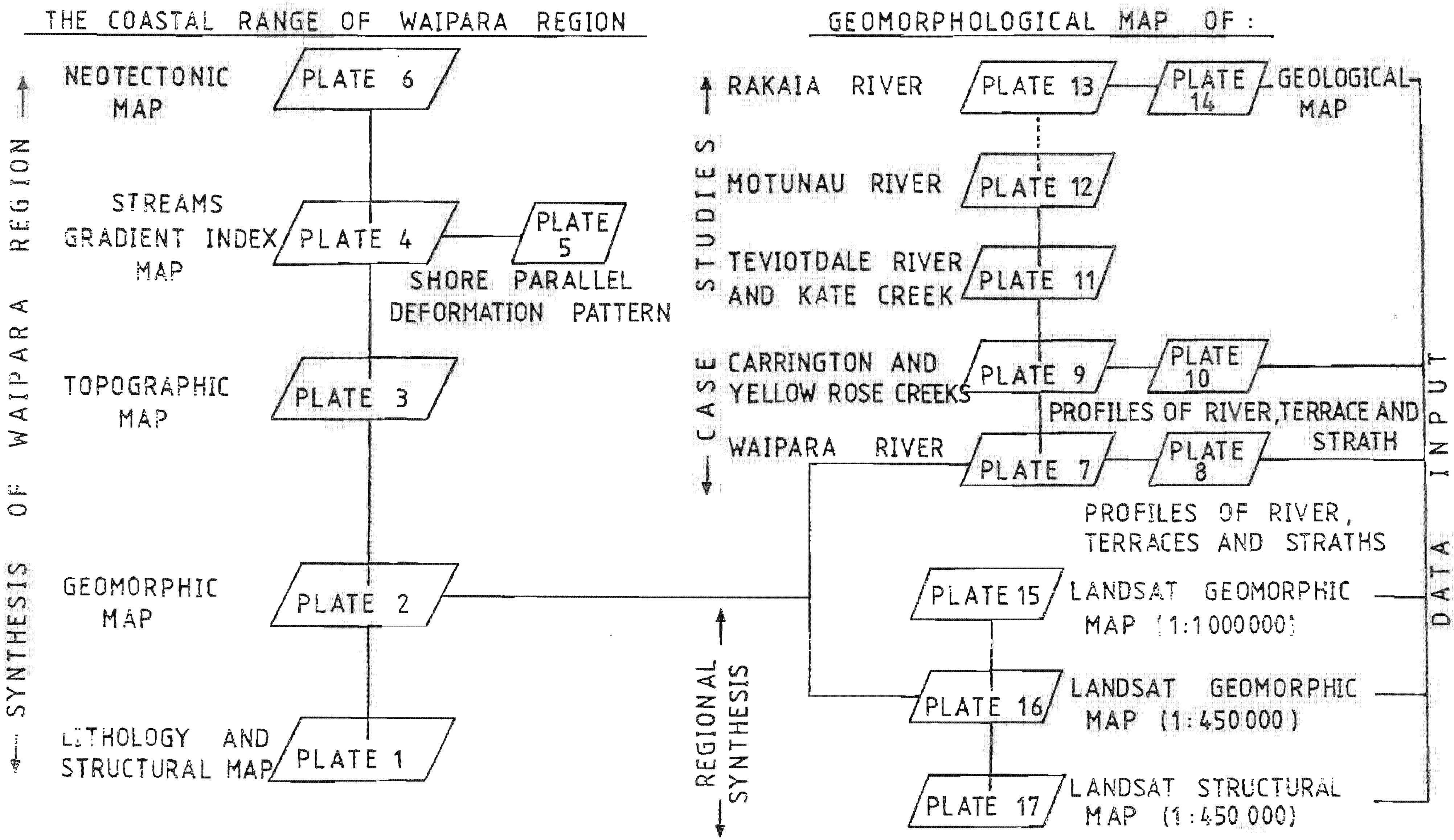
Field work employed standard mapping technique of reconnaissance using traverses and stereoscopic study of aerial photographs, followed by detailed study of all accessible exposures.

Following a preliminary survey, certain critical areas were selected for detailed investigation (Fig. 1.2). These were mapped at larger scales. For remapping, portions of the stereo-aerial photographs were enlarged to scales of up to 1:10,000 and 1:5,000, and a further period of five to six months was spent on field work and ground levelling survey in those areas. Stratigraphy was reexamined in detail and profiles of river, terrace and strath prepared. Wood samples were collected and sent for C14 dating.

1.1.6 CLIMATE

Rainfall is typically of low concentration but of prolonged duration rather than short storms of high intensity. Mean annual

Figure 1-2 Flow diagram showing the connection between component maps and profiles relating subsections of this study



rainfall within the coastal ranges is from about 800-1200 mm (Wards, 1976). The prevailing wind for most of the year is north-easterly with south-westerlies associated with the majority of rain. During winter months (June, July, August) occasional frosts and heavy rains occur in many areas. Mean January temperatures range from about 12 to 22^{o} and mean July temperatures range from 2 to 10^{o}. The climate is broadly classified as humid.

1.1.7 MAPS AND PHOTOGRAPHS

Complete photographic coverage of Waipara area, at several scales and different times is available in the Geology Department/NZGS aerial photograph collection and in the Department of Lands and Survey.

The following aerial photographs were used during this field work. NZ Aerial Mapping Ltd run numbers.

Run 5468 A 10-26)

5468 B 6-21) 1/2/1980

5468 C 10-16) scale (1:25,000)

5468 D 9-13)

Other photographs available but not used for mapping:

Run 1820 - 1831

Published maps of the area used during the investigation are:

- N.Z. Geological Map, 1:63,360, Interim Series NZMS 1, Sheet S68 & S69, Waipara subdivision.
- N.Z. Geological Map, 1:250,000, Sheet 18, Hurunui.
- Base map is available in the Department of Lands and Survey at scale 1:50,000.

1.2 METHODOLOGY

Routinely, field work was preceded by laboratory time spent on air photo analysis and interpretation. This method proved suitable for detailed geological and geomorphological field mapping and was much less time-consuming than more conventional field mapping.

The investigation involved 43 photographs of good quality. Data from the aerial photographs were plotted on the 1:25,000 topographic maps.

The method of presenting information was determined by the availability of topographic maps. The 1:50,000 topographic map was enlarged to a scale of 1:25,000 to match the aerial photographs taken in 1980 at a scale of 1:25,000 and provide the base for photo interpretation. As the original 1:50,000 topographic map was a photogrammetric compilation, adjustments had to be made when adding detail from the distorted photographs to the corrected plot.

The investigation utilised both standard black and white aircraft, and Landsat, prints. Photo-interpretation was carried out using a Topcon stereoscope with the transfer of annotations from the image to the base map being done by handsketching. There were three stages involved:

1.2.1 PRE-OPERATIONAL STAGE

This stage involved careful selection and interpretation of imagery in order to provide a satisfactory foundation for the operational stage.

Firstly, a loose mosaic of aerial photographs was laid on a table to form a composite view of the entire area. Preliminary analysis of air photographs was undertaken using the Topcon stereoscope with higher magnification used on the instrument for the study of details. The interpretation of air photos, using stereo models between the match lines, was carried out.

As a rule the boundaries of mappable lithological and geomorphological units and structural lines could be picked out, but it was more difficult to determine the exact type of rock occurring within the various lithological units distinguished.

1.2.2 OPERATIONAL STAGE

This important phase involved ground data collection and the technique used in transferring information to the base map.

General descriptions of lithology, structure, and geomorphology were obtained during the visits to the field. Geological and geomorphological boundaries were checked and the mapping of forms, processes etc, was done in the field using the lines drawn on the imagery in the preliminary interpretation if they proved to be of use.

The transfer of the interpretation (annotations) from aerial photographs to base map was time consuming. The technique used in transferring thematic information is referred to as a match line using consecutive aerial photographs, with the flight lines as control (Zuidam & Zuidam 1979). The flight lines serve to align the successive overlays in the true flight direction. The photographs were assembled by superimposing detail, and data which had previously been plotted on an overlay were transferred to the base map. The transfer of data by handsketching is a rapid procedure and, although of low precision, is quite acceptable in the case of reconnaissance or even semidetailed surveys. Because of the slight relief, inaccuracies caused by relief distortion were not rectified. Furthermore, no attempt was made to determine distortions caused by tilt. Where detail failed to agree at photographic boundaries, adjustments were made arbitrarily. Locally, these adjustments were considerable. Overall, the base plot control on the adjustments was sufficiently close to preclude the necessity of employing more refined photogrammetric techniques at that scale.

A detailed drainage map of the study area was prepared and used as a base for plotting gradient index data. The contouring, on which the gradient index depends, is from the photo grammetrically compiled base map.

Figure 1.2 shows the synthesis and layers of information in the geomorphological map of the coastal range of the Waipara region.

1.2.3 HIGH RESOLUTION STAGE

This stage was developed after the geological and geomorphological maps were completed (Fig. 1.2), and resulted in several significant and new outputs for the purpose of developing new tools in tectonic geomorphology. For this reason, four target areas were selected in the Waipara area and one chosen for contrast in the dominant modes of deformation outside the study area (i.e. the Rakaia Gorge).

The study involved five phases:

1. A detailed geomorphological map of the selected areas was prepared from aerial photographs on a scale of 1:10,000 and 1:5,000 (approximately). This follows earlier field work supplementing Wilson's basic stratigraphy and structure of the area so that the lithology and structural control on the geomorphology is well understood.

2. Each map was then corroborated and supplemented by field work involving a detailed study of individual geomorphic units and landforms from dynamic, tectonic and morphogenetic aspects.

3. A detailed ground survey using a Wild DI3 Distomat and T1A Theodolite was carried out to prepare profiles of both strath and surface terraces together with the modern flood plain and river bed profiles of the lower Waipara River and Carrington Creek.

4. The geomorphic maps were finalised and different geomorphic units were classified following the scheme devised by the ITC textbook (Zuidam & Zuidam 1979).

5. Taking into account the actively developing and relict geomorphic features, the geomorphic and tectonic history of each study area was reconstructed.

This study is also involved in the application of Landsat imagery to the structural and geomorphological analysis. Further discussion is reserved for Chapter Seven.

1.3 ANALYTICAL METHODS

1.3.1 DIGITIZER ANALYSIS

The coastal range of the Waipara region was analysed by the digitizer device (for more details see Chapter 5). The digitizer is used to feed the co-ordinates of key points directly to an interfaced microcomputer and these data are used in quantitative analysis of topographic characteristics. The tools consist of interactive computer programs which the planner can use without specialised computer training. The equipment belongs to the School of Forestry, University of Canterbury, and I was assisted in its operation by Mr D Mark.

A suite of such programs has been written by Dr B. Hockey, University of Canterbury, Geography Department, with assistance from the writer and Mr D. Mark (see Appendix 1).

1.3.2 SURVEY ANALYSIS

A detailed ground levelling survey using a Wild DI3 Distomat and T1A Theodolite was carried out to prepare profiles of both strath and surface terraces together with the modern flood plain and river bed profiles along the lower Waipara River and Carrington Creek. Appendix 2 shows the collection of the data from each station. The equipment belongs to the Department of Civil Engineering, University of Canterbury, and I was assisted in its operation by Alister McCrae and Gordon McLean.

1.3.3 MAPPING SYSTEMS

Mapping and representation systems of the present sheets are nearly similar to the ITC geomorphological map system (Verstappen & Van Zuidam 1968, Zuidam & Zuidam 1979), and for more details see Chapter Four.

1.4 PREVIOUS WORK

1. The most extensive research previously pursued in the Waipara area was initiated by Jobberns (1937a & b) and was mainly concerned with the geomorphology of the lower Waipara gorge district. The origin of this gorge has been discussed by many authors including Hutton 1905, Cotton 1913, Speight 1912 & 1915, Jobberns 1937a & b, Gregg 1950.

Jobberns (1926, 1928), Jobberns & King (1926) and others were the first who noted the presence of the continuous raised coastal terraces along the North Canterbury coast between the Waipara and Jed Rivers.

2. Gregg (1959) studied in detail the stratigraphy of the lower Waipara gorge, North Canterbury.

3. The geology and stratigraphy are already well documented (Wilson 1963) and the existing geological map is at scale of 1:63,360.

4. An unpublished account of the stratigraphy and chronology of the Hawera Series marginal marine succession of the North Canterbury coast has been described in detail by Carr (1970).

5. Bradshaw & Newman (1979) described low-angle thrusts in Cenozoic rocks in Canterbury, New Zealand.

6. Harris (1982) studied in detail the stratigraphy of the Canterbury Gravels, Omihi River section, North Canterbury.

CHAPTER TWO

STRATIGRAPHY

2.1 INTRODUCTION

This chapter describes the formations present in the study area. The rocks range in age from Mesozoic to Quaternary in the following stratigraphic succession (based largely on Browne & Field 1985):

Oldest

- Torlesse Supergroup	U. Jurassic - L. Cretaceous
- Broken River Formation	Piripauan - Haumurian
- Conway Formation	Haumurian
- Loburn Mudstone	Teurian
- Waipara Greensand	Teurian
- Ashley Mudstone	Teurian - Kaiatan
- Amuri Limestone	Whaingaroan
- Omihi Formation	Waitakain - Duntroonian
- Waikari Formation	Waitakain - Altonian
- Mt Brown Formation	Otaian - Waiauan
- Tokama Siltstone	Otaian - Waiauan
- Greta Formation	Tongaporutuan - Mangapanian
- Greenwood Formation	Wai pipian
- Kowai Gravels	Nukumaruan
- Marine Gravels	Hawera
- Teviotdale Gravels	Waimean
- Canterbury Gravels	Otiran
- River Gravels	Holocene

Youngest

Six figure grid references (such as S68/133073) are based on the national thousand-yard grid of the 1:63,360 topographical map series (NZMS1). Stratigraphic columns are designated in the text in the form S68/C2 (S68 being the 1:63,360 map sheet number and C2 being column number 2).

2.1.1 LITHOLOGICAL SUCCESSION

The geology of the study area (see Plate 1) is of importance from two points of view. Firstly, the lithologies of the area affect the way of which subaerial weathering processes, erosion and deposition have operated. Secondly, those rocks may be eroded and redeposited in the form of river terraces and planation surfaces.

All the deposits known are of sedimentary origin and include terrestrial and marine deposits, the latter laid down under conditions varying from those of the shore-line to upper bathyal depths. A correlation diagram for a coastal northeast transect through north Canterbury describes the known formations in the study area (Fig. 2.1). Particular attention was given in the field to recording the stratigraphic succession by the examination of exposed river sections (Fig. 2.2). The principal differences allowing identification are variable sedimentary, compositional, structure, paleontological evidence, resistance to erosion and consolidation of rock units. The sequence of sedimentary rocks shows that sedimentation was influenced by a change from transgressing to regressing seas. It is not the purpose of the thesis to contribute to stratigraphic refinements or to consider depositional and environmental matters. The primary purpose is to define rock stratigraphic units each of which is distinguished by lithologic characteristics that produce distinctive responses to

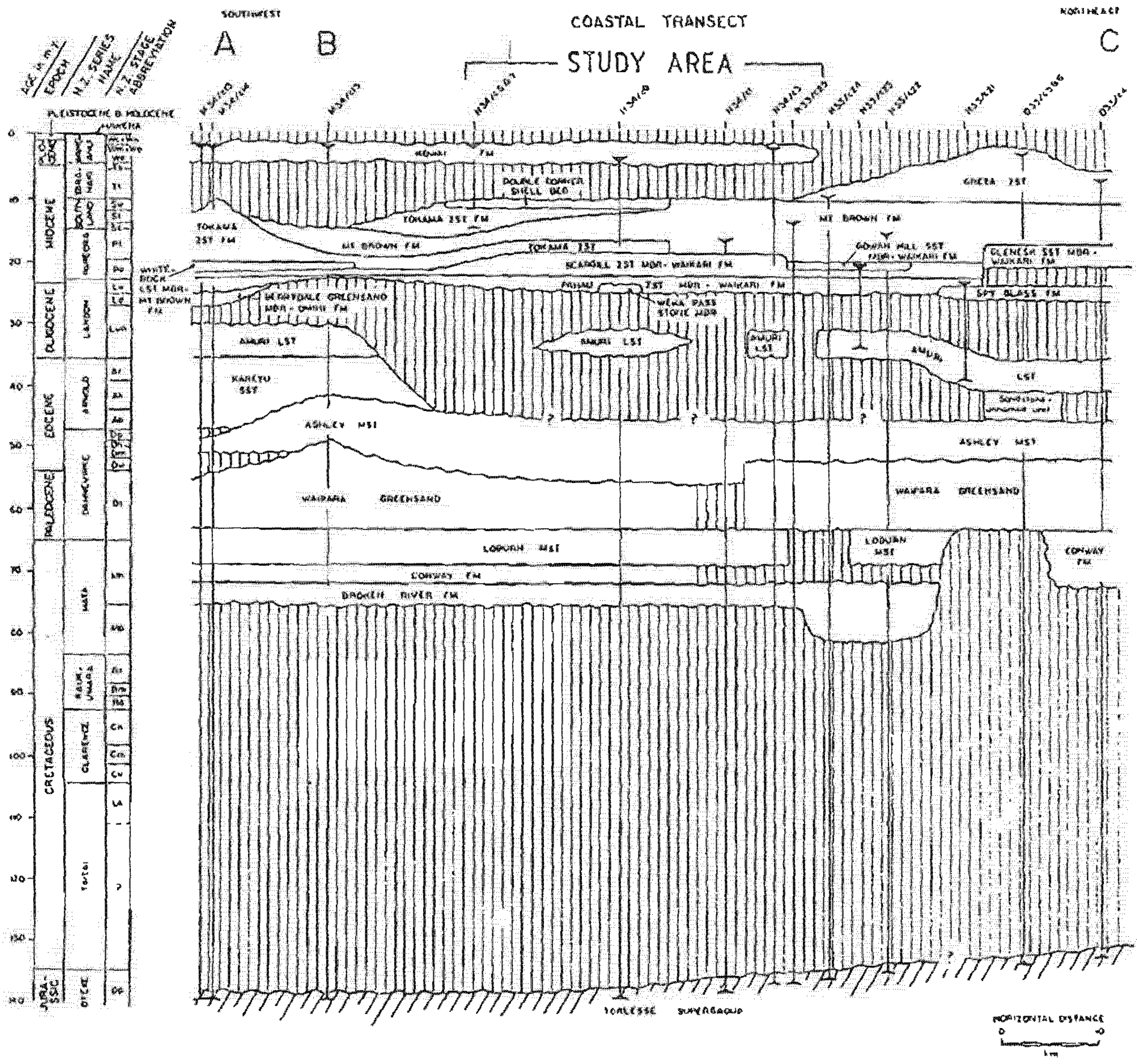


Figure 2.1

Correlation diagram for coastal northeast trending transect through North Canterbury: From A (Mt Grey), to C (Gore Bay) passing through the study area. Vertical line pattern indicates periods of non-deposition or erosion. Vertical bars indicate the interval at which rocks are exposed in each section (after Browne &Field 1985).

erosion and to the development of associated landforms which can be identified from photogeologic mapping. Therefore, descriptions have been kept to generalisations and age relationships and stratigraphic correlations are summaries of existing literature together with studies by the writer. The geological sequence is summaried thus - starting with the oldest formation as follows:

2.2 TORLESSE SUPERGROUP

The Upper Jurassic and Lower Cretaceous Torlesse Supergroup underlies the whole of the study area, and crops out along the cores of anticlinal ridges and in blocks upfaulted in the late Tertiary and early Pleistocene (Plate 1). These rocks are part of Rangitata sequence, a late Carboniferous to Early Cretaceous complex which here constitute the Pahau sub-terrane (Bradshaw et al. 1980).

Throughout the study area, Torlesse Supergroup consists of indurated quartzo-feldspathic sandstone and mudstone, with less abundant/common conglomerate. The structure of the Torlesse Supergroup is rather simple on a regional scale, but proves to be more complex at a smaller scale. Typically the greywackes comprise complexly folded, faulted, and jointed, interbedded mudstone and sandstone, with rare conglomerate horizons. Although the shattering and development of joint planes precludes the accurate determination of dip and strike in many areas, a general north-east south-west trend is apparent from the contact of the greywacke and overlying units.

The Torlesse Supergroup underlies, with pronounced unconformity, much less indurated Late Cretaceous and Tertiary strata.

The Torlesse Supergroup includes a diverse range of lithotypes, the majority detrital. In a recent review of Torlesse rocks (Andrews

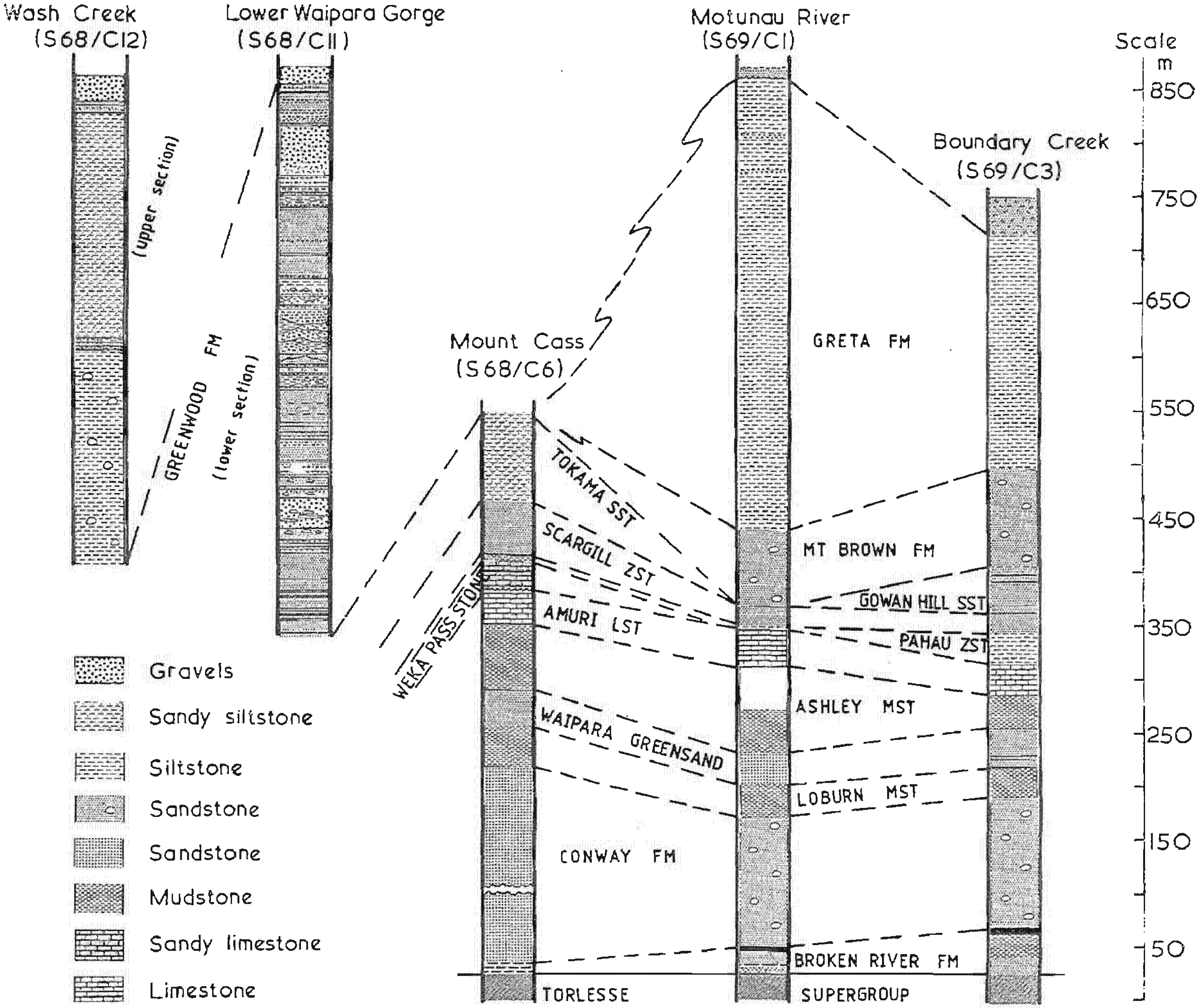


Figure 2.2
Stratigraphic columns illustrating variation in Cenozoic lithological units along the coastal range of the Waipara region.

et al. 1976), depositional environments ranging from non-marine through marginal marine/deltaic to submarine fan were identified.

2.3 EYRE GROUP

The Eyre Sand was defined by MacPherson (1947) and subsequently extended by Andrews et al. (in prep.) to include dominantly sandy rocks at or near the base of the Cretaceous-Cenozoic sequencee in Canterbury. The Group is essentially a transgressive shelf unit of Late Cretaceous to Whaingaroan age.

Seventeen formations are recognized from the Group in North Canterbury by Browne & Field (1985). Only these units that occur in the study area are described here.

Oldest

- Broken River Formation	Piripauan - Haumurian
- Conway Formation	Haumurian
- Loburn Mudstone	Teurian
- Waipara Greensand	Teurian
- Ashley Mudstone	Teurian - Kaiatan

Youngest

2.3.1 BROKEN RIVER FORMATION

(After Gage 1970; modified by Andrews et al. in prep.)

Named from Broken River, western Canterbury, with a type section at Broken River (S66/239962 to S66/241962).

Throughout the study area, Broken River Formation consists predominantly of white, poorly indurated, non-calcareous, well sorted, medium to fine quartzose sandstone and dark grey, poorly indurated, carbonaceous, centimetre to decimetre, bedded, poorly sorted, silt and sand. It occurs widely along the south-eastern limb of Black Anticline, at the northern end of Montserrat Anticline and on the eastern limb and near Montserrat Landslide (Plate 1). The bottom of the unit is defined by a thin localised basal breccia and conglomerate including both greywacke and quartz pebbles up to 2 cm in size. Fine conglomerate lenses occur interbedded with carbonaceous sediment near the top of the formation (S68/239161). The Sandstone shows little evidence of any type of bedding.

Coal beds and coal lenses are more typical towards the base, and are up to 4 m thick. In the study area, good exposures occur at S68/359151. Coalified wood and bark fragments are common, the wood structure being well preserved in stratified tabular siliceous concretions.

Haumurian ages are indicated by micro- and macrofossil evidence for the great majority of the formation. Only at Boundary Creek just to the north of the study area, have Piripauan ages been reported (Browne & Field 1985).

The formation is interpreted as non-marine to marginal marine at the base, becoming fully marine (inner shelf) towards the top. The basal conglomerate facies could be a fluvial deposit left during transgression, while the coal seams are likely to have been deposited in a fluvial flood plain/delta setting. This sequence passes gradually upward into a marginal marine, nearshore lithofacies represented by quartozose sandstone with flaser bedding, and cross stratification. The formation is thickest at Boundary Stream, where 340 m occurs. To the south the formation thins to 140 m at Motunau River and generally thickens again towards the southwest, being 300 m at Mt Cass (Fig.2.2).

2.3.2 CONWAY FORMATION

(modified after Warren & Speden 1978)

Named from Conway River, North Canterbury, with a type section at the north face of Haumuri Bluff (S55/803505). In north Canterbury the Conway Formation consists of light to dark grey, poorly indurated, slightly calcareous, massive, jarositic, bioturbated, slightly glaucontic, poorly to well sorted, silty fine medium sandstone (Warren & Speden 1978).

In the study area the Conway Formation is the major Late Cretaceous (Haumurian) lithology. It is exposed along the flanks of anticlines (Plate 1). Most of the rocks are soft and have been deeply eroded in places by streams which tend to flow parallel to strike. Stream exposures are in many places obscured by slumping. The Conway Formation consists predominantly of sandstone in the lower part, and siltstone and mudstone in the upper; the rocks everywhere showing sulphur efflorescence. Calcareous concretions and rounded pebble sized clasts of sandstone and chert are abundant, more especially in the sandstones. Quartz and minor feldspar are the dominant detrital minerals in the sand and coarser silt fractions. Glauconite is a very minor constituent. Thin bedding is clear in the lower unit (sandstone) but mostly indistinct or absent in the siltstone.

Throughout the study area, Conway Formation unconformably overlies Torlesse Supergroup or conformably succeeds Broken River Formation.

The Conway Formation was deposited in a marine, probably near-shore environment marked by gentle bottom currents and sedimentation from suspension (Warren & Speden 1978). The formation is thickest at its type section, Haumuri Bluff, where 240 m occurs. In the lower Waipara area the formation reaches 100-250 m (Fig. 2.2).

2.3.3 LOBURN MUDSTONE

(after Mason 1941)

Named from the Loburn district, western Canterbury, with a type section at the West Branch, Grey River (S67/882031).

Loburn Mudstone comprises black or dark grey, brown or purple, soft to moderately indurated, calcareous and non-calcareous, jarositic, micaceous, burrowed sandy mudstone (Mason 1941, Strong 1984).

The formation originally covered the whole of the study area but is now preserved mainly on the unfaulted flanks of Tertiary anticlines (Plate 1).

Microfossil evidence indicates a Teurian age (Webb 1966, Strong 1984).

Throughout the study area, Loburn Mudstone grades up from Conway Formation and is succeeded gradationally by Waipara Greensand. The fine grained nature of the sediment, and the gradational change from the underlying Conway Formation suggest enviromental conditions were similar during the time of deposition of both formations, with perhaps less organic matter during deposition of the Loburn Mudstone (Browne & Field 1985). The formation attains a uniform 50-70 m thickness over much of the southern part of its distribution (Figs. 2.1 & 2.2).

2.3.4 WAIPARA GREENSAND

(after Hector 1884, Thomson 1920)

The type area is the Waipara River, from whence the name is derived. The first detailed description was by Thomson (1920) who recorded a lower stratigraphic interval of alternating siliceous, indurated, and poorly indurated greensand beds and an upper, very dark green weakly indurated, argillaceous sandstone. These two units are now formalized as Mt Ellen and Stormont Members respectively (Browne & Field 1985).

On the south-east limb of the Cass Anticline, east of the basement inlier, the more indurated Waipara Greensand forms a step in the high escarpment of the Coastal Range. Elsewhere, Waipara Greensand forms subdued topography and lowlands.

At Cass Anticline, the Waipara Greensand is less than 20 m thick (S68/132218). The upper half consists of decimetre to metre bedded olive-grey, indurated, richly glauconitic, siliceous, moderately sorted, medium to fine, pyritic, silica-cemented sandstone. This forms the prominent escarpment along the south-east limb of the Cass Anticline. The lower half consists of yellow-grey, moderately indurated, siliceous, very glauconitic, moderately sorted, silty mudstone.

On the basis of the foraminiferal evidence Waipara Greensand indicates a Teurian age (Hoskins 1969).

The Waipara Greensand is interpreted as a shallow marine deposit that accumulated under conditions of very slow sedimentation (Browne & Field 1985). In the Waipara area the formation reaches 70-80 m (Fig. 2.2) and thins progressively to the south and only 1.5 m is recorded from Kowai-1 (see Andrews et al. in prep.).

2.3.5 ASHLEY MUDSTONE

(after Mason 1941)

The type section is the West Branch, Grey River as described by

Mason (1941). The formation was named from the Ashley River.

Ashley Mudstone is a blue-grey to medium green-grey, moderately indurated, calcareous, jarositic, bioturbated, bentonitic (especially near the base) mudstone, and glauconitic, well sorted, fine sandy mudstone commonly with pyrite concretions (Browne & Field 1985). The formation becomes progressively less glauconitic in its upper part in the type area. Phosphatic nodules occur locally at the base or near the middle of the formation (Maxwell 1964).

The base of the formation is Teurian to Waipawan (see Hoskins 1969) and the top is dated as Kaiatan by Foraminifera.

Throughout the study area, Ashley Mudstone rests conformably, or disconformably (with a burrowed contact), on Waipara Greensand. It is conformably overlain by Amuri Limestone. At the type section 35 m is recorded and is 75 m thick at the mid Waipara section (Fig. 2.1).

Slumping due to the bentonitic nature almost everywhere obscures Ashley Mudstone beds.

2.4 AMURI LIMESTONE

(modified after Hutton 1874)

This distinctive white or pale grey, argillaceous limestone always in thin beds of Haumurian to Early Oligocene age is widely distributed throughout Marlborough and North Canterbury and first described and defined by Hutton (1874). The Amuri Limestone is defined as centimetre to decimetre bedded (stylobedded), light cream (weathered colour) to green-grey (unweathered), indurated, bioturbated (*Zoophycos* is particularly common - see Lewis 1970) calcilutite: coccolith and foraminiferal, slightly glauconitic biomicrite (Browne & Field 1985).

In the study area, the greater part of the limestone is a very hard, chalky white, fine-grained rock, closely jointed at the surface into small cuboidal blocks. At the base it passes rapidly into a glauconitic limestone and then into creamy marls which grade down into very glauconitic calcareous mudstones (S68/286119).

Throughout the study area, Amuri Limestone rests conformably upon Eyre Group. The top of the formation is planar, but in detail the contact may be locally irregular with infillings of glauconitic sands of overlying Motunau Group sediments.

In the study area compositional changes are slight in the Amuri Limestone, the thickness being a few metres to 80 metres (Fig. 2.1). This is in part due either to post-depositional faulting and removal by intra-Oligocene erosion or has occurred as a result of one or more intra-Oligocene erosion periods , perhaps related to a drop in sea-level (see Lewis & Belliss 1984).

The formation is diachronous being oldest in coastal Marlborough, and becoming progressively younger toward the south. In the study area the formation is Whaingaroan in age (Browne & Field 1985).

2.5 MOTUNAU GROUP

(Browne & Field 1985)

Motunau Group is widespread in North Canterbury. The Group is dominated by sedimentary rocks and coarse bioclastic limestone. The Group includes the Omihi Formation of Late Oligocene age to the Kowai Formation of Pliocene to early Pleistocene age (Browne & Field 1985). It excludes the late Quaternary outwash gravels and tills. The Group marks a period of largely shallow marine sedimentation that culminated in Late Miocene to Pleistocene regression. Only six formations occur in the study area (Fig. 2.2):

Oldest	
- Omihi Formation	Waitakian - Duntroonian
- Waikari Formation	Waitakian - Altonian
- Mt Brown Formation	Otaian - Waiauan
- Tokama Siltstone	Otaian - Waiauan
- Greta Formation	Tongaporutuan - Mangapanian
- Kowai Gravels	Nukumaruan
Youngest	

2.5.1 OMIHI FORMATION

(after Andrews 1963)

The Omihi Formation is widely distributed throughout the area. The formation consists of five members - only one of which is relevant to this work, the Weka Pass Stone.

The type section is the Weka Pass railway cutting, (S61/087225) as described by Andrews (1963). The member was named by Hutton (1877).

The Weka Pass Stone is typically a thoroughly cemented, cream to light grey, massive, unjointed, sandy limestone with fine disseminated glauconite. The member is usually thickly bedded with some borings throughout, mainly near the base. At the base cobbles of phosphatised Amuri Limestone are locally common (Andrews 1963).

The Weka Pass Stone forms distinctive escarpments throughout the study area. These escarpments are more resistant and generally easy to distinguish from the underlying Amuri Limestone by its unjointed massive appearance, except near the base, where it is softer owing to a concentration of glauconite .

In the following localities Weka Pass Stone is absent.

1. A few kilmetres north of the Motunau River on the eastern limb of Montserrat Anticline. The prominent limestone escarpment in that area being formed of Amuri Limestone (S68/380162).
2. Northern end of Montserrat Anticline, on the western limb (S68/312156).
3. Along the south-eastern limb of Black Anticline (S68/250156).

In the study area the Weka Pass Stone is disconformable with the underlying Amuri Limestone. The contact is bored, the borings filled with glauconitic sandstone. The time interval represented by this disconformity or disconformities has been the subject of recent investigation work (Findlay 1980, Carter et al. 1982, Carter 1985). At some sections, there is a gradual transition between Amuri Limestone and Weka Pass Stone (Andrews 1963). The Weka Pass Stone conformably underlies Waikari Formation.

Foraminifera indicate the Weka Pass Stone to be largely Waitakian in age. At some places the base of the member is Whaingaroan and at others Duntroonian. Macrofossil evidence is meagre, but tends to support these determinations. Wilson (1963) gives a full list of the rather fragmentary shells obtained from the Weka Pass Stone.

In the study area Weka Pass Stone varies in thickness from a few metres to 60 metres. It is thickest in the south but thins progressively to the north (see Figs. 2.1 & 2.2).

2.5.2 WAIKARI FORMATION

(after Andrews 1963)

This formation includes the blue-grey and grey sandstone and siltstone that overlie the Omihi Formation and underlie the Mount Brown Formation, throughout North Canterbury (Andrews 1963).

The formation consists of five members - only three of which are relevant to this work, the Pahau Siltstone, the Scargill Siltstone, and to the east the Gowan Hill Sandstone. Two of Andrew's diagrams are here included (Figs. 2.3 & 2.4) showing the areal and stratigraphic distribution of the three respective members.

Throughout the study area, the more indurated sandstones frequently form prominent escarpments. The less indurated sandstones and siltstones of the Waikari Formation erode to form topographic depressions between the indurated sandstones and also between the Weka Pass Stone which is stratigraphically below of the Omihi Formation. This alternation of indurated and less indurated tilted beds produce a distinctive monoclinal structure over much of the area wherever the Waikari Formation occurs.

On the basis of foraminifera evidence the Waikari Formation shows an extreme age range of Waitakian to Altonian. Most of the formation is Otaian, but in some sections the top is as young as Altonian (Andrews 1963).

The Waikari Formation conformably overlies the Omihi Formation throughout most of the area. However to the north-east it disconformably overlies Amuri Limestone, as at Motunau River (Fig. 2.2).

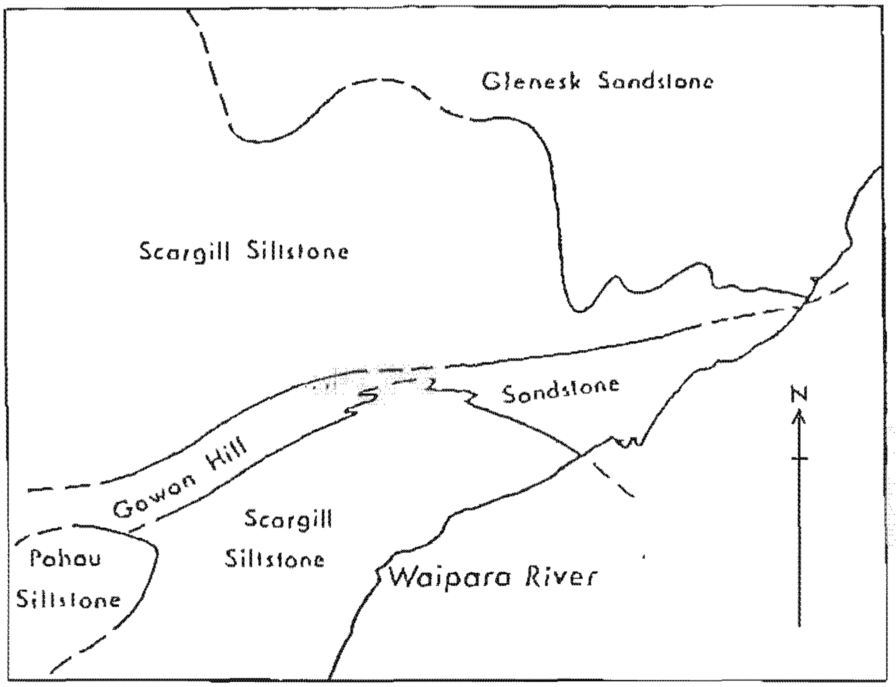


Figure 2.3
Andrew's areal distribution of members of the Waikari Formaton (after Andrews 1968).

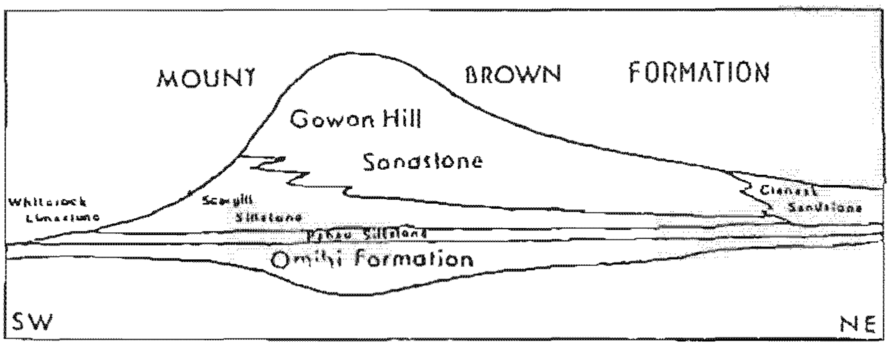


Figure 2.4
Andrew's stratigraphic distribution of members of the Waikari Formation in relation to the Mount Brown Formation (after Andrews 1963).

A. Pahau Siltstone (Member)

(modified after Speight & Wild 1918; Andrews 1963)

The name is taken from the Pahau River into which flows Cascade Stream, the type section at (S54/059508).

Pahau Siltstone is typically a grey-green, massive, highly glauconitic, calcareous sandy siltstone (Andrews 1968). The distinguishing lithologic features of the Pahau Siltstone are a moderate glauconite content, and the sandy-silt grain size. It is everywhere sandy at the base where it overlies the Weka Pass Stone, and rapidly becomes silty and less glauconitic towards the top. Where the Pahau Siltstone overlies Amuri Formation, the base is a nodular layer containing cobbles of phosphatised white limestone and large grains of glauconite resting on the eroded surface of the limestone (at Motunau River). The Pahau Siltstone is about 2 m thick (Fig. 2.2).

Throughout most of North Canterbury the Pahau Siltstone conformably overlies, and shows a transitional contact with, the Weka Pass Stone (Andrews 1963). In a few sections the Pahau Siltstone disconformably overlies Amuri Formation (at Motunau River ,S69/C1).

Foraminifera indicate a Waitakian-Otaian age for the Pahau Siltstone throughout the study area (Andrews 1963).

B. Scargill Siltstone (Member)

(modified after Mason 1949; Andrews 1963)

The name and type section is taken from Scargill Creek (S61/266305).

The Scargill Siltstone is a light blue-grey, partly indurated, calcareous sandy siltstone containing a very small amount of fine glauconite and mica. The sediment is massive, although indistinct coarse bedding is commonly revealed by differential weathering (Andrews 1968).

At Mount Cass escarpment the member is composed of light blue-grey, massive, siltstone at the base. Higher up it consists of bands of light brown-grey, slightly sandy, faintly glauconitic siltstone alternating with bands of the typically light blue-grey, calcareous siltstone. Towards the top, the siltstone is more clay rich, with only a few bands of harder, brown-grey, sandy siltstone. The contact between these bands of fine and coarse siltstone are everywhere gradational.

In the study area, the indurated sandy siltstone forms prominent escarpments. The less indurated sands and silts erode to form topographic depressions.

Rich assemblages of Foraminifera show the Scargill Siltstone to be Otaian at most places. An exception is the Mount Cass escarpment where the base of the member is Otaian and the top Altonian (Andrews 1963).

The Scargill Siltstone conformably overlies the Pahau Siltstone. The member is 103 m thick at the Mount Cass escarpment. Elsewhere it has an average thickness of 30-60 m thickness (Fig. 2.2).

C. Gowan Hill Sandstone (Member)

(after Andrews 1963)

The type section is 6 Km south-west of Gowan Hill Homestead (S68/050185) as described by Andrews (1963).

The member comprises a sequence of thick bands of brown-grey, calcareous, fine sandstone and thinner bands of light blue-grey, calcareous siltstone (Andrews 1963). Calcareous tabular concretions which occur at the base of the Gowan Hill Sandstone, emphasise stratification; elsewhere the unit is unstratified.

In the study area, the Gowan Hill Sandstone is about 20 m thick, and consists of alternating thick beds of yellow-brown, indurated, massive, calcareous, well sorted, fine sandstone and light blue-grey, calcareous siltstone. The siltstone beds decrease regularly in thickness toward the top of the unit (at Motunau River). At some localities, calcareous tabular concretions may occur within a thick sandstone unit. They are about 15 cm thick and persist along the strike (such as at S68/305162).

Foraminifera show the age of the Gowan Hill Sandstone to be Otaian in most sections. At Motunau River the top metre is Altonian (Andrews 1963).

Throughout the study area, Gowan Hill Sandstone conformably overlies the Scargill Siltstone. The lower contact is sharp, and is marked by a basal layer of thoroughly bored, brown-grey, glauconitic, muddy siltstone (Andrews 1968). The maximum thickness is 190 m at Middle Waipara River, while at Motunau River 20 m is present (Fig. 2.2).

2.5.3 MT BROWN FORMATION

(modified after Haast 1871b)

The formation is named from Mt Brown (S68/991107), south of the Waipara River.

Mt Brown Formation consists of a diverse range of lithotypes including limestone, conglomerate, sandstone and siltstone lying stratigraphically above Waikari Formation (Browne & Field 1985). In the study area, the Mt Brown Formation is dominated by medium brown (weathering to a light grey-brown), soft to moderately indurated, calcareous and non-calcareous, massive to moderately well laminated, micaceous, lithic, well sorted medium to very fine sandstone. Limestone and conglomerate units are not exposed in the study area, but inland it rests conformably on Waikari Formation.

The formation is well exposed at Motunau River (S69/C1), where it is 70 m thick and rests conformably on Waikari Formation. In most places, the indurated sandstone forms prominent escarpments. The less indurated sandstone and siltstone erode to form topographic depressions.

The formation is widely distributed throughout the study area. It is conformably overlain by Greta Formation at Motunau River, interdigitates with the Tokama Siltstone at the lower Waipara area, or is disconformably overlain by the Greenwood Formation or Quaternary gravels (Fig. 2.2).

Foraminiferal and macrofossil determinations indicate Otaian to Waiauan ages (Browne and Field 1985).

2.5.4 TOKAMA SILTSTONE

(modified after Mason 1941)

The West Branch of Grey River (S67/886033) was designated the type section by Andrews et al. (in prep.).

In the lower Waipara area the Tokama siltstone consists predominantly of blue-grey, moderately indurated, calcareous, fine sandy siltstone with scattered shell fragments and calcareous concretions. The top of the formation is marked by two fossiliferous pebble conglomerate bands separated by blue-grey sandstone, collectively known as the Double Corner Shell Bed (Gregg 1959, Bradshaw & Newman 1979).

In the study area the blue-grey calcarous Tokama Siltstone is restricted to the area between the lower Waipara Gorge, Limestone Creek, and Dovedale Stream (Fig.2.2). Most of the siltstones are soft and have been deeply eroded to form topographic depressions but the more indurated sandstones frequently form prominent escarpments.

Throughout the study area, Tokama siltstone overlies Scargill Siltstone apparently conformably. The top of the Double Corner Shell Bed is in fault (thrust) contact with the overlying Greenwood Formation in the lower Waipara River as suggested by (Bradshaw & Newman 1979) who recently advanced a theory of low angle thrust tectonics in the region. Further discussion is reserved for Chapter three.

An age range from Otaian to Waiauan is indicated by Foraminifera in the samples from the Grey River and Amberley areas (Browne & Field 1985).

The Tokama Siltstone is likely to represent a basin of reduced sediment input, deposited perhaps in middle to outer shelf conditions. The Double Corner Shell Bed is a coarse-grained conglomerate facies, which marks the filling of the Tokama basin (Browne & Field 1985).

2.5.5 GRETA FORMATION

(modified after Hutton 1888)

The type section is designated in Cobbolds Creek-Greta River (S62/401292 to S62/420291) Browne & Field (1985).

The formation does not occur widely in the study area. It is largely confined to the southeastern flank of Montserrat Anticline that extends from Motunau River to at least Hurunui River in the north outside the study area. The formation is well displayed at Motunau River and Boundary Creek (S69/C1&3), where it is 400 m thick and rests conformably on Mt Brown Formation.

At Motunau River the formation consists of blue-grey, moderately indurated, thin-bedded calcareous mudstone-siltstone and muddy fine sandstone (Rogers 1970, Lewis 1976).

Most of the formation is Tongaporutuan to Mangapanian. At Motunau, the base extends down to the Lillburnian (Browne & Field 1985).

Maxwell (1964) believed the formation was deposited in quiet water, probably on the continental shelf or slope. The thickness of the formation implies a continually subsiding basin. Molluscs indicate upper bathyal conditions.

2.5.6 GREENWOOD FORMATION

(after Gregg 1959)

The name Greenwood Formation is here continued for the marine gravel, conglomerate, sand, siltstone and mudstone which crop out at the lower Waipara area as proposed by Gregg (1959). The best section is exposed on both sides of the Waipara River near Greenwood's Bridge (S68/C11). The beds exposed at the left bank of the Waipara opposite its junction with Omihi Stream, belong to the upper part of the Greenwood Formation (S68/C12).

At the base of the Greenwood Formation there are everywhere present, several bands of hard pebble and shell conglomerates separated by beds of fine grained sands. The conglomerate consists of well rounded greywacke clast about 0.5 to 2.5 cm in diameter. Shells are abundant but mainly fragmentary. The conglomerate bands are prominent ridge-forming beds of the study area.

The upper part of the Greenwood Formation is characterized by the presence of a considerable thickness of cemented shell beds, and greywacke pebble conglomerate and gravel, interbedded with sands. Conglomerate becomes progressively more abundant upsequence. Decimetre to metre-bedded, light yellow-brown, moderately indurated, massive, slightly calcarous, fine sandy mudstone forms a less abundant lithotype. Fossils are abundant but are mainly fragmentary and difficult to extract due to the degree of weathering. Many shells are preserved only as casts or moulds.

Throughout the study area, Greenwood Formation disconformably overlies older Tokama Siltstone. Elsewhere it unconformably overlies older formations.

The age of Greenwood Formation was determined to be Waipipian (Dr A.G. Beu personal communication to S. Nathan 1979) due to the presence of Mesopeplum and Chlamys (Phialopecten) triphooki marwicki (Beu).

2.5.7 KOWAI GRAVELS

(after Gregg 1959)

Speight (1919) proposed the name Kowai Gravels for the tilted nonmarine sands and shellbeds overlain by rapidly alternating marine gravel, sand, and silt of the Waipara River area. Thomson (1920) restricted the name Kowhai (sic) Series to the non-marine gravel. Gregg (1959) used the name Kowai Gravels and restricted them in the lower Waipara Gorge to the non-marine gravels.

Gregg (1959) defined the division between the Greenwood Formation and overlying Kowai Gravels as being:

"where sediments containing shells and shell fragments pass up into sands and gravel bearing no trace of marine origin", (Gregg 1959, p.522).

The concept that the Greenwood Formation is marine and is overlain by nonmarine Kowai Gravels (as used by Gregg 1959) is adopted in this study. The description of Gregg (1959) is more appropriate to the study area, hence the nomenclature of Browne & Field is not followed here.

Throughout the study area Kowai Gravels are best preserved in the lower Waipara Gorge (S68/134034), left bank of Omihi Stream (S68/134138) , and south of Weka Pass. The formation is predominantly gravel, but there are lenses and beds of sand and silt. Near the top of the formation beds of lignite occur. Erosional surfaces precede each gravel unit. The gravels are generally massive, moderately weathered, poorly sorted gravel conglomerates with rounded to well rounded Torlesse sandstone cobbles and rare silty-clay rip-up- clasts. The matrix is a sandy silt. Sometimes a lime-rich silt was exposed towards the top of the sequence.

At most places the Kowai Gravels are succeeded by Hawera Series (late Quaternary) outwash gravels and associated sediments, with pronounced unconformity, and grades downwards into the Greenwood Formation.

The microflora from the lignite of the Kowai Gravels in the Greenwood bridge section suggests a Nukumaruan age (Gregg 1959). The Kowai Gravels are probably a deposit of fluviatile origin with possible lacustrine and estuarine phases.

2.6 MARINE GRAVELS

Thin marine deposits (the Hawera Series high-level marine gravels of Carr 1970, Fig. 2.5), rest with angular unconformity upon Tertiary rocks and consist of shelly gravels and sands, the gravels containing flat pebbles principally of Torlesse-derived rocks, but with minor Tertiary pebbles.

Throughout the study area, this marine terrace can be traced south from Boundary Creek to near the Waipara River, with a reduction in height toward the south (see plate 1).

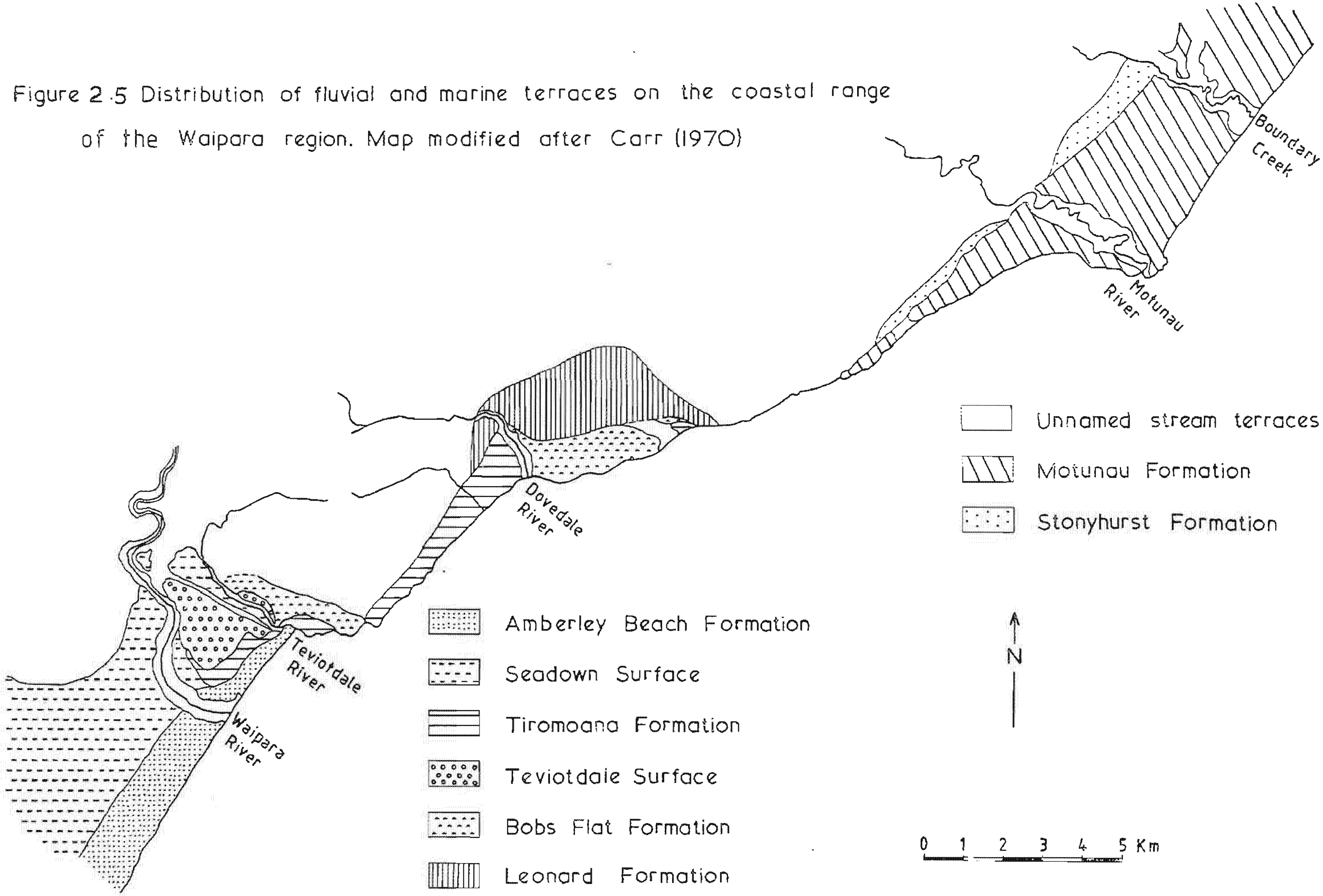
Previous workers in North Canterbury tended to concentrate on lithostratigraphic nomenclature for geographically restricted areas. This has resulted in a proliferation of stratigraphic names as in the case of the marine terraces along much of the coastal range in the Waipara region. For example Carr (1970) defined six coastal formations near the coast of Hawera Series age along the 33.5 km of coastal range from Waipara River to Boundary Creek (see Fig.2.5). Carr suggested an age range from Waiwheran to Aranuian. The description of these formations is retained here, but in reality only five levels of marine terraces are adopted in this study (see Table 2.1). They are, from youngest to oldest: Postglacial Marine Terrace (M), 60 ky Marine Terrace (M1), 80 ky Marine Terrace (M2), 105 ky Marine Terrace (M3), 125 ky Marine Terrace (M4). Major Pleistocene marine terraces in uplifted areas can represent high sea levels in successive interglacial stages. Further discussion is reserved for Chapter five.

2.6.1 STONYHURST FORMATION

(modified after Speight & Wild 1918; Carr 1970)

The Stonyhurst coastal plain comprises high (146-204 m) remnants

Figure 2.5 Distribution of fluvial and marine terraces on the coastal range of the Waipara region. Map modified after Carr (1970)



Carr's Nomenclature		This study		
Motunau sector	Dovedale and Teviotdale sectors	Stage	Name	Age
Stonyhurst Formation	Leonard Formation	Waiwheran	Marine Terrace M4	125 ky
			Marine Terrace M3	105 ky
	Bob's Flat Formation	Terangian	Marine Terrace M2	80 ky
Motunau Formation	Teviotdale Gravels	Waimean	Teviotdale Gravels	?70-75 ky
	Tiromoana Formation	Oturian	Marine Terrace M1	60 ky
Unnamed River Gravels	Seadown Formation	Otiran	Canterbury Gravels	
	Amberley Formation	Aranuian	Marine Terrace M0	Holocene

Table 2: Correlations of proposed nomenclature with that of Carr's for the late Quaternary deposits of the coastal range of the Waipara region.

landward of the Motunau shoreline (Fig.2.5). Within the mapped area, the coastal plain comprises two dissected remnants that persist as a narrow step fringing the Montserrat Anticline on both sides of Motunau River (see plate 1).

The Stonyhurst Formation consists of two terrace surfaces based on their heights and geomorphic evidence. Two former shorelines are well preserved north of the Motunau River, but poorly preserved south of the river and correspond to the last interglacial high sea levels at 105 and 125 ky respectively, for this reason the Stonyhurst Formation is subdivided into two marine terraces.

Carr (1970, p.118) gave a full description of the type locality (S62/516238), where about 3.3 m of well-stratified greywacke gravel rests with angular unconformity upon a Tertiary basement surface. The gravels are moderately yellow-weathered, though contain little clay matrix. Individual pebbles are all discoid in shape. Shell grit constitutes a 5 cm layer at the base of the section. Pebble morphology, gravel texture and stratification, and fossil content, show that the sequence is a beach deposit.

Stonyhurst Formation is assigned to the Waiwheran Stage (Carr 1970).

2.6.2 LEONARD FORMATION

(modified after Jobberns 1926; Carr 1970)

The Leonard coastal plain consists of little more than a collection of flat-topped spurs and related topographic accordances deeply eroded by tributaries of Dovedale Stream and coastal creeks (Plate 1). The terrace remnants are extensively preserved north of Dovedale Stream, and are subdivided into two levels based on their heights and geomorphic positions. The upper marine terrace is well preserved on the coast in the range from 216 to 228 m a.s.l., and its inner margin is bordered by distinct sea cliff (125 ky shoreline) which run concave to the present day shoreline. The lower marine terrace (105 ky) is preserved on the coast about 201 m a.s.l. The outer edge is defined by a nearly vertical escarpment 76 to 91 m high. This scarp represents a younger sea cliff of 80 ky ago.

Carr (1970, p.29) gave a full description of the type locality (S68/276105), where there is a 0.91 m thick exposure of gravel. It is composed of well sorted, well stratified, greywacke pebble gravel. The pebbles average 5 cm in diameter, and are highly rounded and discoid. It rests with angular unconformity on all older rocks.

Leonard Formation is tentatively referred to the Waiwheran (Carr 1970).

2.6.3 BOBS FLAT FORMATION

(modified after Jobbern 1926; Carr 1970)

The Bobs Flat coastal plain comprises two dissected remnants (Fig. 2.5) about 5.63 km apart characterized by an irregular hummocky surface (Carr 1970). Firstly, at Dovedale Stream the coastal plain remnants lie between the lower Leonard coastal plain (105 ky terrace) and a series of slumps parallel to the present day coastline. To the south the more deeply eroded portion of Bobs Flat coastal plain can be subdivided into three levels at a few localities as shown in plate 1. The higer terrace surface is bordered at its inner margin by steep a slope, which is considered to represent an old sea cliff (125 ky), capped by slope wash deposits. The lower terrace (80 ky) is separated from the higer terrace remnant by a distinct riser that is apparently an old sea cliff. At a few places it is subdivided into two levels of low risers. The heights of former shorelines at the backs of the terrace range from 128 -146 m a.s.l. at the higher terrace level and from 100 - 121 m a.s.l. at the lower terrace level.

Carr (1970, p.33) gave a full description of the type locality (S68/264089). The sequence consists of 1.3 m of well stratified fine greywacke gravel overlain by 1.5 m of light brown silt. The gravel, well cemented by iron oxide, is composed of very well sorted greywacke pebbles up to 1.2 cm diameter, and is regarded as a beach gravel. The overlying consolidated silts are identical to contemporary slopewash deposits derived from nearby hills of weathered Tertiary rocks.

Bobs Flat Formation rests unconformably on a smooth surface cut on Tertiary siltstone. A Terangian age is suggested by Carr (1970). This age agrees with that of Gregg (1964), who correlated Bobs Flat Formation with the Parikawa Formation of South Marlborough.

2.6.4 MOTUNAU FORMATION

(modified after Jobberns & King 1933; Carr 1970)

Motunau Coastal Plain is that planar landform along the eastern slopes of the Montserrat Anticline from near Montserrat landslide to at least Stonyhurst Creek in the north outside the study area. Within the mapped area, the Motunau coastal plain is widest (3.2 km) near the Motunau River mouth, and progressively narrows towards the north and south. The plain forms a narrow coastal step of about 228 m wide near the Montserrat landslide (Platel). The Motunau coastal plain is bordered at its inner margin by a conspicuous former sea cliff. The former shoreline on the terrace is well developed in the range from 112 to 124 m a.s.l., although the surface is now completly covered by surficial deposits of alluvial fan and slope wash. The Motunau coastal plain is inferred to be the 80 ky marine terrace.

Carr (1970, p.125) gave a full description of the type locality (S69/416178), where about 14.6 m of well stratified greywacke gravel rests with a smooth-cut surface upon Tertiary blue-grey siltstones. The sequence consists of 5.4 m of very well-stratified seaward-dipping greywacke gravel which forms the basal and thickest individual member of the formation. Pebbles within each stratum are particularly well-sorted and fall within the size range 0.63 - 2.54 cm. The pebbles are very well-rounded and discoid. 0.9 m of poorly stratified muddy very fine sand rests upon the greywacke gravel member, the change from gravel to sand being everywhere abrupt. Up-section the muddy sand passes into yellowish red, very muddy fine gravel, consisting of angular greywacke chips and well rounded flattened greywacke pebbles, akin to those in the basal stratified gravel member. Stratification of the gravel is indistinct.

About 2.0 m of non-stratified, very fine sand and mud, succeeds the mixed greywacke gravel member. Whereas its lower contact is fairly sharp, the top 0.6 - 0.9 m of sand includes numerous greywacke flakes and chips up to 10 cm in diameter. Thus the up-section change back to gravel is strictly gradational. This second angular greywacke gravel unit shows most of the characters of the first, but is almost totally lacking in well rounded pebbles. Again, the gravel is very muddy and shows only poorly developed stratification. Completing the reference section is about 3.65 m of structureless yellow-brown flaky mud, barren of both pebbles and fossils.

All aspects (stratification, pebble sorting and shape) of the basal stratified greywacke gravel member suggest deposition in a beach environment, a conclusion that is supported by the occurrence of a marine platform of boulders commonly bored by marine molluscs. The overlying sediment succeeding the beach type gravel was deposited under alluvial conditions; the muddy matrix and the angularity of the pebbles indicates fan (colluviul) or stream deposition.

The major portion of Motunau Formation is assigned a (?) Terangian-Waimean age and is correlated with both Bobs Flat Formation of Terangian age, and of Oturian-Otiran age at the mouth of Motunau River and is correlated with Tiromoana Formations (Carr 1970).

2.6.5 TIROMOANA FORMATION

(modified after Jobberns 1926; Carr 1970)

This is a narrow coastal plain that persists continuosly for 10.45 km from north of the Waipara River mouth to Dovedale Stream (Fig.2.5). The last interglacial shoreline on the terrace is well develop in the range from 100 to 38 m a.s.l. The outer edge is defined by a nearly vertical coastal cliff 34 to 65 m high. The Tiromoana coastal plain is inferred to be the 60 ky marine terrace.

Carr (1970, p.47) gave a full description of the type locality (S68/233080). The sequence consists of medium grey, moderately sorted, medium sand, in a bed (0.82 m) thick resting upon the cut platform. This sand unit generally shows horizontal thin bedding, with some small-scale trough cross-stratification. Small well rounded greywacke pebbles up to 2.5 cm diameter are scattered throughout the sand, and shell material is not uncommon. Overlying the sand unit is a dominantly clayey-section some 1.8 m thick. It can be divided into three units. At the base is a 2.1 cm blue-brown mottled clay, rather sandy at its top and bottom. The clay contains a freshwater fauna. The upper and lower contacts of the unit are gradational, the lower part being interstratified with grey sand. 12.7 cm of unfossiliferous orange iron-stained medium sand rests upon the clay. Upon the sand, with gradational contact, rests 0.85 m of poorly stratified blue clay, with small tree stumps and carbonaceous fragments near the top. Overlying the blue clay unit a 10 cm thick lens of silty gravel rests with erosional discordance. Pebbles that comprise the gravel are entirely of Tertiary lithologies, dominantly limestone. Succeeding the limestone gravel is a thick (4.7 m) stratified grey-brown mud. A steep erosional surface (about 30 ^{o}) separates grey-brown mud from the overlying unit, 4.8 m of very coarse sand, gravel, and white silty clay. Massive structureless yellow-brown mud, about 4.8 m thick, completes the section. It lacks stratification, is highly calcareous, flaky, which weathers to give a vertically fluted surface.

In general terms, the reference section is predominantly of non-marine deposits. Clearly the basal grey sand unit has marine affinities. An environmental change to freshwater conditions is indicated at the first appearance of blue-brown mottled clay. Whereas the blue carbonaceous clay is thought to be a freshwater (lagoonal) deposit, the overlying succession is regarded as terrestrial (Carr 1970).

Almost horizontal strata of the Tiromoana Formation rest with angular discordance (about 15^{o}) on tilted late Tertiary blue-grey siltstone. The Tiromoana Formation is referred to the Oturian Stage (Carr 1970).

2.6.6 AMBERLEY BEACH FORMATION

(Carr 1970)

The Amberley Beach coastal plain is the northeast extension of a flat strip of land lying between the present -day shoreline and the Amberley "postglacial" sea-cliff (Fig. 2.5). Within the mapped area, the Amberley Beach coastal plain is widest in the mouth of Waipara River, and progressively narrows towards the mouth of Teviotdale Stream, to less than 0.7 km (Plate 1).

Carr (1970, p.113) informally used the term "Amberley Beach Formation" to include all deposits of the coastal plain. Dominant are beach sands and shingle, that constitute a series of subparallel ridges southwest of the Waipara River. Fluvial greywacke gravels of the contemporary river floodplain, as well as swamp muds and dune sands, are included within the formation, as are fan gravels and silts that overlie beach deposits of the landward edge.

Gregg's (1964) map shows the Amberley Beach Formation and its extensions as Holocene. In terms of Suggate's (1965) revised Quatermary time-scale, the Formation would be of Aranuian Age.

2.7 TEVIOTDALE GRAVELS

(modified after Jobberns 1937a)

Jobberns (1937a, p.131) named the Teviotdale Terrace (here called Teviotdale Gravels) and described it as being best developed on the north side of the Waipara River, extending from the gorge to near the present day shore.

Wilson's (1963) informal proposal of the term "Teviotdale gravels" for thick river deposits surficially represented by the "Teviotdale Surface" included notes on content, relations to older formations, and conditions of deposition.

Carr (1970, p.41) gave a full description and erected a type locality (S68/140050), where about 21.3 m of moderately well stratified gravel rests with marked unconformity upon all older rocks. The gravel is a very poorly sorted with rare pods and lenses of finer material, mostly very fine sand and silt. Pebbles are composed almost entirely of greywacke and are subangular to rounded. The sequence is yellow-brown, a reflection of the weathered state of the pebbles, which are soft enough to be easily shattered with a hammer.

The gravels have been deposited during a period of fluviatile aggradation (Gregg 1959).

No carbonaceous deposits that might date the Teviotdale gravel have been found. The gravels are much younger than Nukumaruan and their deposition probably accompanied the penultimate major advance of the last glaciation (Wilson 1963). Carr (1970) follows Suggate in referring the Teviotdale Gravel to the Waimean Stage.

2.8 CANTERBURY GRAVELS

(after Wilson 1955)

The Canterbury Gravels were named by Wilson (1955) and underlie what he termed the "Canterbury Surface". In the study area, the surface of these sediments dips gently to the east to meet the low hills at the west side of the coastal range (Plate 1).

Wilson (1955) formally described the Canterbury Gravels as including;

"the whole of the material deposited by aggrading rivers during the cycle of aggradation that culminated in the formation of the Canterbury Plains." (Wilson 1955, p.127).

and also

"the aggradational gravels that underlie equivalent surfaces in North Canterbury, even where these are not apparently glacial outwash, e.g. deposits adjoining the Waipara River." (Wilson 1955, p.128).

Wilson (1955) included flat-lying sediments in the Waipara -Omihi Valley in this definition. Harris (1982) recognised seven depositional facies within the Waipara -Omihi valley of either fluvial or slope-formation origin. A meandering channel facies model is proposed for the deposition of the fluvial sediments.

The term Canterbury Gravels is adopted in this report to include both flat-lying sediments in the Waipara-Omihi valley and Canterbury Surface.

Wilson (1955) correlated the Canterbury Surface with the last major advance of the Otira or Last Glaciation. The absence of periglacial deposits (e.g. loess and solifluction) on later surfaces supported Wilson's correlation.

Harris (1982) correlated the sediments from the Canterbury Surface in the Waipara-Omihi valley, although not glacial outwash, with sediment deposited during and since the Poulter advance of the Otira Glaciation. A single radiocarbon date was obtained from a Hyridella sample collected from a shellbed in the uppermost silt unit (S68/ 926942). The date obtained was 10,550 ± 150 years. This date corresponds well with the predicted age of these sediments. Detailed discussion related to this date is however postponed to the chapter five.

2.9 RIVER GRAVELS

In the study region the geomorphic terrace surface can be divided into three groups based on elevation above stream level. First, the modern floodplain, which ranges from 1 to 2.5 m above the water level is the most laterally persistent and youngest depositional feature. Second, a low terrace at heights from 2.5 to 35 m above the modern floodplain is defined by a group of terrace remnants of diverse lithology and age. Third, a high terrace is represented by only a few isolated remnants at heights various from 35 to 60 m above the modern stream. These latter remnants are the oldest river deposits recognized in the study area. They consist predominantly of greywacke gravels with very minor amounts of Tertiary rocks. Plate 1 shows the distribution of river terraces and floodplain within the study area. Detailed discussion related to these terraces is however postponed to the Chapter six.

CHAPTER THREE

STRUCTURE

3.1 INTRODUCTION

Previous studies (Speight 1912, Smart 1954, Gregg 1959, Wilson 1963, Carr 1970) have demarcated anticlinal and synclinal structures by using topography, dips and strikes, structure contour mapping and time-stratigraphic unit boundaries, or some combination of these. In the present study, emphasis has been placed on the use of aerial photographs to delineate these anticlines and synclines. Basically the fold pattern has been modified by the fault system which has caused portions of the folds to be faulted out with only remnants being preserved. The strongly deformed nature of the cover rocks is clearly demonstrated by the disturbance and swing of the strike ridges throughout the study area.

In the present study, the structure of the area has sometimes been found to differ from earlier interpretations. This structure may be summarised as consisting of a group of folds and faults arranged in a crude en echelon pattern which is strongly reflected in the resulting outcrop pattern.

3.1.1 REGIONAL DEPOSITIONAL AND STRUCTURAL SETTING

Three major depositional and two major structural phases are recognisable within the North Canterbury Region. Although the structure of the study area (Fig. 3.1) is not complex in comparison with some other areas in New Zealand it has, nevertheless, several interesting structural features whose development is summarised below.

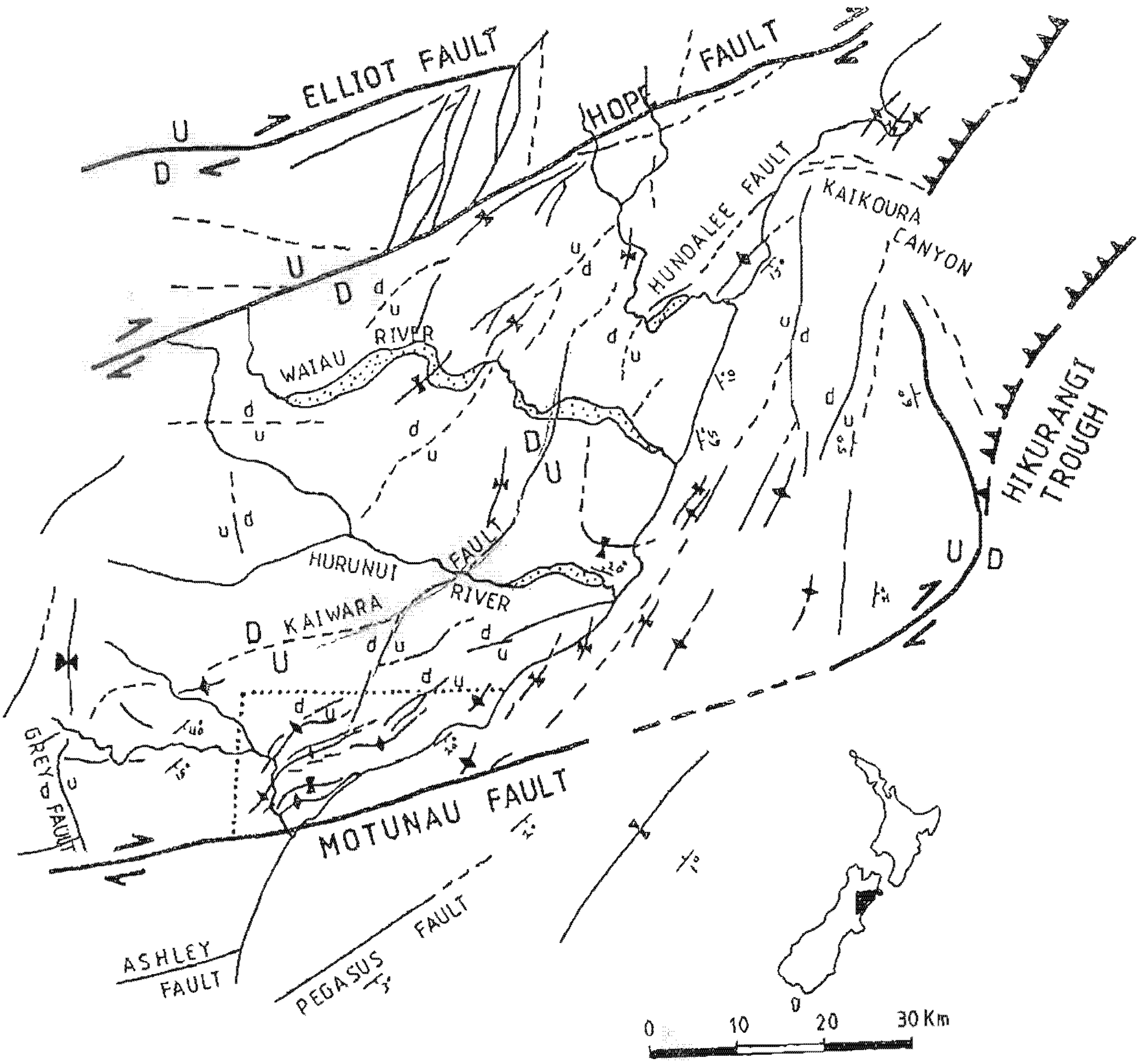


Figure 3.1

The setting of the study area in the North Canterbury- South Marlborough Region. Onland data from Gregg (1964), offshore data from Carter & Carter (1982). The study area is bounded by dotted lines. Insert - The location of this area within the South Island, New Zealand.

The present structural framework of the region was initiated during the Mesozoic following the deposition of the Torlesse Supergroup. This local basement is a thick clastic succession now recognised as part of the Torlesse terrain that was complexly deformed during the late Mesozoic Rangitata Orogeny (Bradshaw 1980, Andrews et al. 1976) and now separated from late Cretaceous-Cenozoic strata by marked angular unconformity. These Upper Jurassic and Lower Cretaceous rocks are presently exposed in the cores of anticlinal ridges and in blocks upfaulted in the late Tertiary and early Pleistocene.

The second period of deposition formed the cover sequence which consists of a marine transgression-regression sequence. Contrasts in the lithology of deposits laid down between Late Cretaceous and Early Pleistocene suggest that the basin has been intermittently uplifted (Browne & Field 1985). The contact between Amuri Limestone and Weka Pass Stone marks the end of the earlier trangression, and the beginning of regression. The Cenozoic cover-sequence has been disrupted by tectonic activity which reached a climax in the late Pliocene and early Pleistocene. This activity is usually referred to as the Kaikoura Orogeny and, within the area, is manifest as a series of en echelon folds with axes generally trending from east-west to east-northeast (Fig. 3.1). The folds are asymmetric and have anticlines whose northwest limbs are either steep or inverted. Normal, reverse, and strike-slip faults, parallel to or forming planes slightly arcuate to the relative bedding, are commonly developed in the sedimentary cover. Reverse faults predominate and have displacements seldom greater than 1200m (Wilson 1963).

The third and latest depositional phase involves the unconformable deposition of sediments on the older strata during Upper Pleistocene to Recent times. In common with many areas in the northern part of the South Island, the region has been intermittently uplifted throughout the Quaternary. Faulting and folding has continued until the present time as the Recent sediments are, in places, deformed (Campbell & Yousif 1985).

3.1.2 REGIONAL STRUCTURE

There are a number of abrupt changes in the style of deformation within the study area (Fig. 3.1). The Cenozoic cover-sequence of the coastal range is complexly deformed by faulting and tight folding, whilst similar sequences to the south are gently dipping; steeply tilted and folded sequences also occur north and west of the study area. That the same contrast in tectonic activity persists to the present day, is shown by the folding and faulting of high level marine strandlines of Pleistocene age which border the eastern margin of the study area (Carr 1970). At the same time parts of the Canterbury Plains have undergone net subsidence (Wilson 1963). For example, observations on the northwest limb of the Black Anticline (Fig. 3.2) suggest that faulting is now the dominant mode of deformation whilst folding has not produced any noticeable effect on the Canterbury Gravels in the Omihi valley (Harris 1982).

Fig. 3.1 gives the location of the study area in the northeast of the South Island. The Hikurangi Trough, a filled trench (Katz 1974), is situated just east of the volcanic, seismic, sedimentological and structural features of the North Island subduction complex (Lewis 1980). The major fault system (Marlborough

Shear Zone) is composed of anastamosing subparallel faults, all of which have the same direction of offset splay from the Alpine Fault. These faults have developed since about 15 million years B.P. to accommodate an increase in obliquity of the direction of movement of the Pacific plate relative to the Alpine Fault plate boundary. In all probability, as these have split apart, regional movement along them has lead to either convergence, compression and uplift, or, to divergence, extension and subsidence. These zones of tension and compression have been called pull-apart basins by Crowell (1974). The Marlborough Shear Zone is not well understood but, although most writers agree that the Marlborough Faults are dominantly strike-slip, opinion is divided as to how the faults were initiated. There is still uncertainty as to whether all faults of the system are equally active or whether they have developed sequentially to the south in response to changes in plate motions (Scholz et al. 1973).

The coastal fold belt is characterized by mainly asymmetric and northwest-verging faulted anticlines, many of which have basement exposures at their core (Gregg 1964).

The complex nature of the fold structures in North Canterbury, many of which are non-cylindrical and often influenced by an intricate pattern of basement faulting, has been further discussed by Bradshaw (1975). The cause of faulting in the rigid basement is possibly due to a shear zone (Rynn & Scholz 1978) that extends from the Hikurangi Trench to the Alpine Fault.

On the basis of first motion studies of the Marlborough region, the depth of the Benioff zone, and the bathymetric trends of its trench, Arabasz & Robinson (1976) concluded that the Hikurangi Trench is migrating southwards. Rynn & Scholz (1978) give further support to

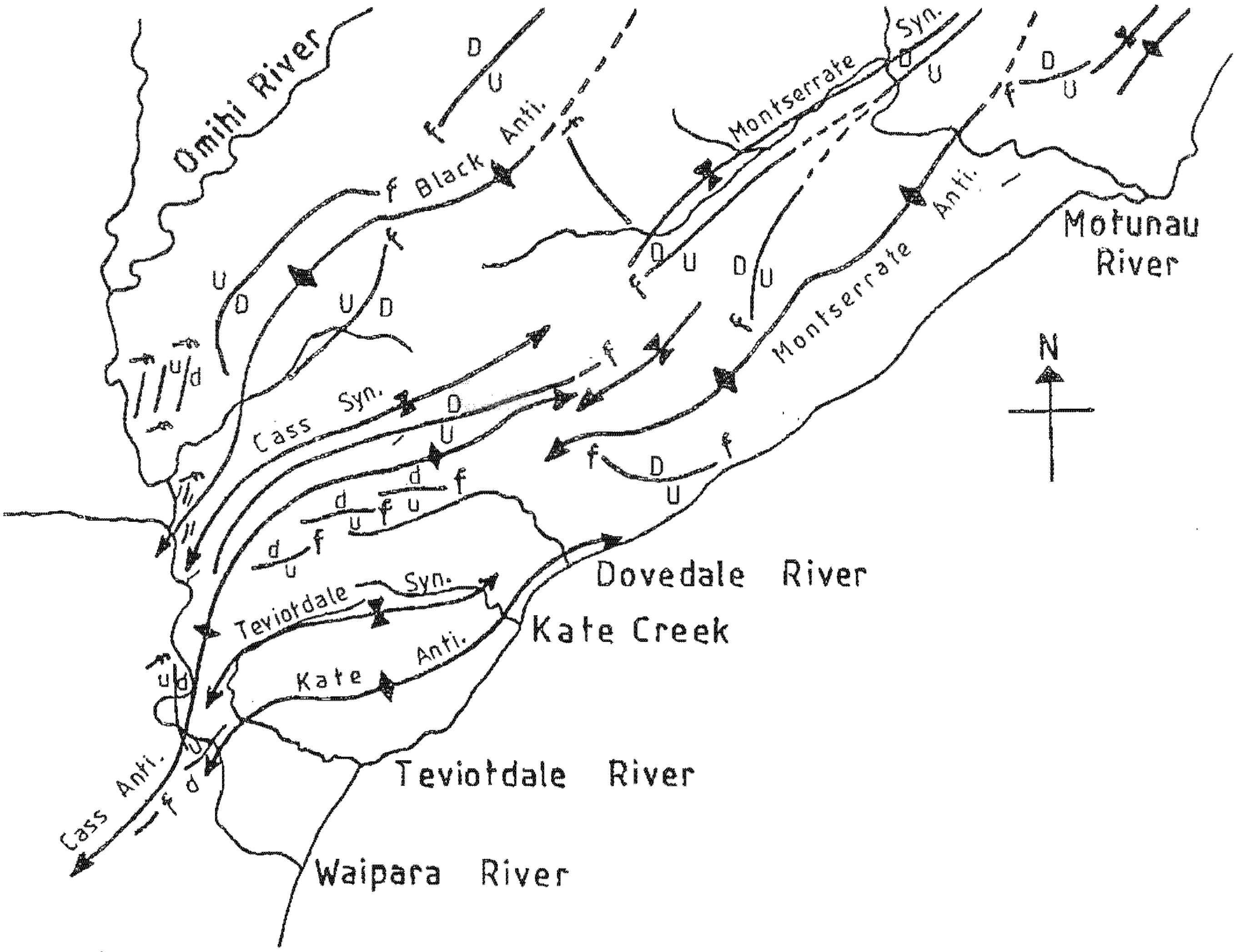


Figure 3.2

A simplified structural map of the study area.

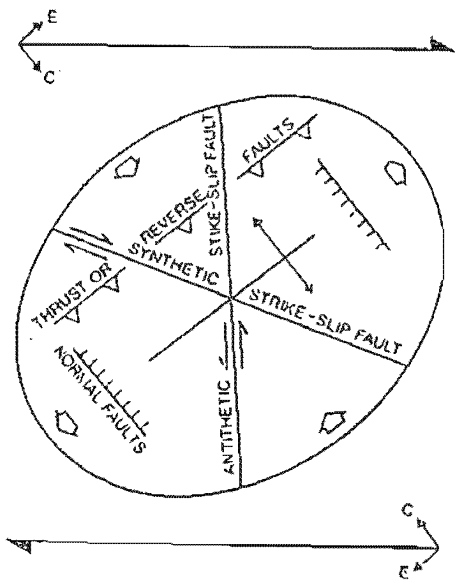


Figure 3.3

Structural pattern resulting from simple shear and produced by a SW - NE dextral shear couple (after Harding 1974).

this model, noting the abundance of seismic activity south of the Hope Fault. They consider that shear deformation is now taking place south of the Hope Fault. These highly seismic settings of overall lateral motion may, at any one place, have very rapid differential vertical movements sometimes exceeding the strike-slip motion. These authors further suggest that the southern boundary of the shear zone, as indicated by the model, is about 50 km south of, and parallel to, the Hope Fault, which is precisely the location for the Motunau Fault System suggested by Carter & Carter (1982).

Carter & Carter (1982) concluded that the Marlborough Shear Zone developed in the early-middle Miocene as a transform fault system linking the oblique-slip Alpine Fault with the subduction boundary along the Hikurangi margin. The Hope Fault is the southernmost well developed fault, although the Porters Pass Fault and the offshore Motunau Fault may represent parts of a zone of deformation that could develop into another major strike-slip fault. Faults of the shear zone have developed successively southwards, with the Motunau Fault System and the Pegasus Bay Fault being the most recently formed (Fig. 3.1).

Tectonically, the North Canterbury-South Marlborough region can be considered to be one in which a transference of motion between the Alpine Fault and the Hikurangi Trench is occurring. The fault and fold structure is complex due to interference and the merging in this region of major tectonic elements, such as the east coast fold belt and the Hikurangi Trough with its tectonically active apex. The area to the west of the study area is influenced by the abrupt, fault-controlled uplift of the Southern Alps (Gregg 1964). Here, in the Mt Grey-Lees Valley district there is a pronounced belt of superimposed, north-south trending fractures and faults. To the north of the study area lies a transform fault system, the Marlborough Shear Zone, where the existence of pull-apart basins has been postulated along the Hope Fault by (Clayton 1966, Freund 1971). Thus the study area lies on the southern boundary of the Marlborough east coast fault system, and extends northward to include several pull-apart basins such as the Hanmer Depression along the Hope Fault. If these authors are correct in their model, one might expect to find that the tempo of tectonic activity has been accelerating up until the present time.

3.1.3 MECHANISM OF FOLDING

Tectonic events have produced a sequence of anticlines and synclines whose structural control of the geomorphology has been fundamental. Generally, the folds are arranged in a crude en echelon pattern, with scarps typically being developed on the harder limestones of the Weka Pass Stone, the Waikari Formation, and the Greenwood Formation, and with axes trending from east-west to northeast-southwest.

Deformation is considered to be the result of a mechanism of progressive en echelon folding. Bedding attitudes were recorded, where possible, over the whole area (Plate 1) and are consistent with a single phase of folding. Along the western margin of the Black Anticline, a decrease in dip with increasing distance from the fold axis was shown. A possible model to explain this type of structure would be deposition accompanied by deformation. The axial traces of en echelon folds are discontinuous when traced through the whole fold system (Fig. 3.2).

En echelon folds are valuable indicators in cover rocks of the presence of strike-slip faults at depth, of the sense of the movement and of the existence of sedimentary pull-apart basins below the cover rocks. Movement is caused by a zone of plastic shear deep within the basement which generates the shortening force in the upper basement rock and in the overlying sediments, leading to characteristic secondary faulting and folding (Harding 1974, Graham 1978, Ponce de Leon & Choukroune 1980, Rodgers 1980, Robertson & Woodcock 1980, Berthe & Burn 1980). It is postulated, from a number of lines of evidence, that the structure of the area can be interpreted in terms of a SW-NE dextral shear regime.

Fig. 3.3 illustrates the development of en echelon folds associated with dextral shearing of the basement. The en echelon arrangement, which has counterparts elsewhere in the world, requires a right lateral couple across a roughly southwest trend, and the northeast-southwest trend of the individual anticlines developes from the associated NW-SE compression. This right lateral couple has existed along a substantial length of the coastal fold zone, from at least the late Pliocene to the late Pleistocene. The maximum compression and extension axes of these folds are comparable to the minor and major strain ellipse axes of the model. With increasing strike-slip displacements, evolutionary changes in pattern may be anticipated and, depending on whether the response of the deformed terrain is dominated by faulting or folding , or by a mixture of both, the specific deformation types may be predicted. Fold orientations in a shear zone can be different for several folds along the same fault trend. Some folds, or parts of folds with irregular axial trends, may parallel the shear fault or cross the shear zone at a low angle

(Wilcox et al. 1973). Several factors that can influence the shape and trend of en echelon folds include convergence of blocks during shearing, changes in strike of the shear fault, a large component of vertical displacement, differences in kind and thickness of sediments, and mobility of the basement near the folds. On the basis of field mapping, most major folds in the area trend northeast-southwest. The northwestern limbs of many anticlines are disrupted by high angle reverse faults; some of the anticlines and faults are arranged in southwest-trending dextral en echelon systems (Plate 1).

Several pieces of evidence suggest that some faults on the flanks of the Black Anticline were tectonically active in the late Pleistocene (for more details see section 3.5.2). These faults form an en echelon system that trends southwest. The parallelism of the faults to each other and to anticlines farther east, and the similarity of their dextral en echelon arrangement to that of the anticlines and their attendant reverse faults, suggests a genetic relation to regional structures. A hypothesis compatable on a regional scale is that the individual northeast trending scarps in the low hills to the east of the Omihi River (S68/138134) generate reverse faults caused by NW-SE compression, and that the southwest trend of the dextral en echelon pattern is due to a right-lateral couple across a southwest-trending structural zone that includes most of the folds in this region.

For the eastern side of the study area, Carter & Carter (1982) concluded that tectonic activity north of the Motunau Fault has continued through the late Pleistocene as shown by the presence of growing folds on the sea-floor. Similar features do not occur to the south (Fig. 3.1).

For the south-western side of the study area, Marden (1976) has revealed evidence of a shear zone bordered along its western side by the Porter's Pass Fault and along its eastern margin by a fault trace, for which abundant evidence is available. These define a zone of active dextral shearing that shows signs of having been active during the late Pleistocene.

The study area is centered on the dominant structures of the Cass Anticline and the Montserrat Anticline (the "Cass Anticline" of Wilson 1963). This folding occurred throughout the Pleistocene since high marine strandlines of Pleistocene age along the eastern margin of the area have been deformed by faulting and folding, (Carr 1970). Thus, the regional tectonism of dextral shearing in the basement along the strike-slip faults has resulted in the dual-plunging nature and the en echelon pattern of many of the folds throughout the study area (Plate 1). This involved deformation of the sedimentary cover as displacement occured along the underlying shear zone in the basement, which was accompanied by a developing pattern of faults which broke through the overlying cover with the increase in the accumulated shear strain. The folds rotate with increasing angular shear giving rise to the sigmoidal pattern and the width of the folded belt also increases with time.

3.1.4 CONCLUSIONS

This study area lies along the southernmost dextral strike-slip fault system of the Marlborough Shear Zone which developed in the early-middle Miocene. The development of en echelon folding, througout the study area, which is associated with SW-NE dextral shear regime from the late Pliocene to the late Pleistocene, suggests that faults of the shear zone have developed progressively southwards. It is possible that this area is in the early stages of the developement of a pull-apart basin, especially as the tectonic activity is less vigorous now than in the Miocene. The observed deformation indicates that most of the faults and folds are arranged in a southwest trending, dextral, en echelon system.

The structural features within this system share similar characteristics of formation and deformation i.e en echelon reverse faulting and folding. These qualities are ideal for testing the efficacy of aerial photographs and landsat images as tools in developing a new approach to the problems of morphotectonics.

3.2 PREVIOUS CONTRIBUTIONS TO THE STRUCTURAL ANALYSIS OF THE WAIPARA AREA

3.2.1 COMMENT ON THE STRUCTURE CONTOUR MAP OF WILSON (1963)

The structure contour map drawn by Wilson (1963) shows the present day three dimensional form of the folded sequence. This map is based on time-stratigraphic unit boundaries and shows the structural contours of the cover sequence.

Throughout the study area, the contour surfaces indicate that the structure has been formed during one phase of folding. The folds are assymetric, with anticlines whose northwest limbs are either steep or inverted. Reverse faults, a common development in these anticlines, together with local overturning are defined by the discontinuity of the structural contours.

Structural contours on the base of the cover sequence are confined to the centre of the anticlinal ridges while structural contours on the base of the Pliocene (Greenwood Formation) extend around the area, and along the subsequent en echelon fold pattern.

3.2.2 PREFOLD THRUSTING AND DECOLLEMENT

Bradshaw & Newman (1979) described low-angle thrust faults from coastal North Canterbury. They inferred that thrusting predated the main phase of faulting and folding, and attributed it to synsedimentary sliding. Katz (1982) on the basis of subsurface data, concluded that the dome-shaped Kowai Anticline, the "Cass Anticline" of Wilson (1963), was an asymmetric, disharmonic fold that verged to the northwest, in which some thrusting and decollement may have occured within the Pliocene. Work by Carter & Carter (1982) has revealed a probable, low-angle of thrusting on a 3.5 kHz profile near the south end of the Conway Ridge, where it too predates the tilting of the affected sediments. Results of the structure contour work undertaken by Wilson (1963), have not produced any evidence of thrusting. Characteristics of the fold forms are not indicative of polyphase deformation.

The results of the author's levelling survey at the Horseshoe Bend (S68/134077) suggest that the thrust fault, described by Bradshaw & Newman (1979), is probably a steep reverse fault. Definite evidence for this is available from recent river terrace deformation along the fault trace. This deformation has been produced by repeated fault movements, the significance of which is discussed in Chapter Six.

3.3 PRINCIPAL ANTICLINES

An overall fold pattern (Plate 1) for the study area can readily be obtained by the measurement of dips and strikes from which the essential anticlinal and synclinal features can be constructed. These are described in sequence from East to West.

3.3.1 KATE ANTICLINE

This anticline was first defined by Wilson (1963). It is an east-striking fold south of the Mount Cass Road. The fold is asymmetrical; on the northern flank the dip varies from 09°-29°NW, whereas the southern flank dips less steeply at around 18°S. The axial trend is roughly northeast-southwest, but lack of exposure prevents detailed measurement. The axial trace, as defined in this study, appears roughly to coincide with Wilson's (1963) axis, but is plotted herein as intersecting the coast further to the east i.e. east of the Dovedale Stream mouth area, as shown on Plate 1.

In general this anticline forms a topographic high. As with fold style, drainage patterns developed on both limbs of the anticline also indicate asymmetry. The gently dipping limb shows a long parallel to subparallel drainage pattern while the steep limb shows a short one. This anticline visibly affects only upper Tertiary sediments with the basal conglomerate of the Greenwood Formation serving as a key marker in defining the outcrop.

According to Carr (1970), the youngest established movement on the Kate Anticline appears to be Oturian (-?Otiran). Tilting is recognisable on the four uplifted marine terraces - the Leonard coastal plain (M4), the Bobs Flat coastal plain (M2), the Teviotdale Surface and the Tiromoana coastal plain (M1).

3.3.2 CASS ANTICLINE

One of the main structural features of the study area is the Cass Anticline first recognized by Speight (1912, p.227-30). Wilson

(1963) interpreted the Cass Anticline as a NE-trending anticline extending northwards from the Makerikeri River, to a point beyond the northeastern boundary of this study area. In reality, Wilson's Cass Anticline consists of three separate en echelon folds. The term "Cass Anticline" is retained for the middle of the three folds. The more northerly segment which also occurs in the study area is renamed as the Montserrat Anticline and is discussed in Section 3.3.4.

The fold is asymmetrical. The beds on the south-east limb dip from 12^{o}-27^{o}SE, those on the north-west dip steeply and are in places overturned. The position of the fold axis is well defined in the lower Waipara gorge and it plunges to the southwest towards the Amberley Hills. Further to the north, the fold axis is seen generally to correspond closely to the trace of the Hamilton Fault and finally it plunges towards the east (Plate 1). The flanks of this anticline are to some extent disrupted by faults parallel to the anticlinal axis (see section 3.5.1).

In Wilson's map, the Cass Anticline was linked to a similar structure in the Montserrat area. From the results of the present study, it seems probable that the fold axis dies out close to the end of the Hamilton Fault (Plate 1). The lack of exposure prevented detailed mapping and the exact location of the plunge.

According to Carr (1970), Wilson's (1963) Cass Anticline appears to have been active throughout the early Quaternary and until at least as late as Waimean times.

3.3.3 BOUNDARY ANTICLINE

This gentle anticline was shown but unnamed on Wilson's map (1963). It is herein named after Boundary Creek.

In this central part of the Motunau coastal plain (M2) Carr (1970) recognised two anticlines. The first, on the inland coastal plain, near Boundary Creek, warped the shore platform and cover strata. The second at the seaward edge of the plain, was considered to extend from south of the Motunau River almost to Boundary Creek, with tilting of the Motunau estuarine beds. Carr (1970) suggested the youngest established movement on the first anticline was Terangian (?Waimean) in age, whilst the latest movement on the second anticline was considered to have been Oturian (?Otiran).

Neither of the two anticlines is evident on the aerial photographs due to the uneven, dissected, low relief. Exposures are restricted to the banks of the incised drainage valleys. My own dip measurements, whilst defining the Boundary Anticline fail to identify the existence of the Motunau Anticline as proposed by Carr (1970); this will be discussed in Chapter Six. The height-age distribution of marine terraces confirms the Late Quaternary growth of the Boundary Anticline.

The fold is asymmetrical, the north western limb having dips that are generally higher than those on the south eastern limb. The axial trend is roughly east-northeast west-southwest, but lack of exposure precluded detailed measurement. It is roughly coincident with the more northerly of Carr's (1970) two axial traces.

3.3.4 MONTSERRAT ANTICLINE

This is an east-northeast trending anticline extending from the Montserrat trig. to beyond the northeastern boundary of the study area. The name is taken from the Montserrat Homestead. This fold is considered to be a separate structure from Wilson's Cass Anticline

(Wilson 1963), being the north-eastern member of the set of three en echelon folds previously mentioned.

The fold is asymmetrical. Beds on the southeastern limb dip from 20°-58°S; those on the northwestern limb dip steeply and are, in places, overturned. The position of the fold axis is well defined at the Montserrat trig. Further west a southerly plunge is developed. Traced northeastwards there is a remarkable swing in the trend of the fold axis to give a sinuous shape caused by the disturbance of the Glendhu Faults (S68/326152) on the northwestern limb.

The Montserrat Anticline is responsible for part of the seaward tilt of the Motunau coastal plain (M2) and was active throughout the early Quaternary and until at least as late as Waimean times (Carr 1970).

There is also some disturbance due to faulting on the southeastern flanks of this anticline (see sections 3.5.5 & 3.5.7).

3.3.5 BLACK ANTICLINE

This anticline was first recognised by Speight (1912, p.227). Wilson (1963) described the Black Anticline as a north east trending anticline extending from the Waipara River to the Black Hills, Scargill. He noted that there was a well defined sag in the axis that produced two domes.

The Black Anticline is distinctly asymmetrical; the beds on the southeast limb dip from 19°-32°SE; those on the northwest dip steeply and are in places overturned. At several locations on the northwestern limb higher dips are recorded but these can generally be explained by local disturbances due to the effect of faults or large scale slumping (see section 3.5.2). The change in strike of the axis of the Black Anticline as well as the position of the fold axis is seen generally to coincide with Wilson's (1963) axis. There are however marked differences in the axial trend at the south end of this anticline. On the basis of my own field data, the fold axis is now located to the west and extended further to the south to intersect the Waipara River (Plate 1). The flanks of this anticline are to some extent disrupted by faults which cross and are parallel to, the anticlinal axis and according to the result of the present study, many new faults are recognised near the plunges of the southern dome (see section 3.5.2).

3.4 PRINCIPAL SYNCLINES

3.4.1 TEVIOTDALE SYNCLINE

This syncline was shown but unnamed on Wilson's map (1963). It separates the Cass Anticline to the north from the Kate Anticline to the south. The name is taken from the Teviotdale Homestead.

The fold is asymmetrical. Beds on the southern limb dip from 09^{o}-20^{o}NW, those on the northern limb dip from 15^{o}-22^{o}SE.

In general this syncline occupies a topographic low and is readily defined by outcrops of Greenwood Formation on either side of the upper reaches of the Teviotdale River. Here, the course of the river is coincident with the synclinal axis with the beds on the northern side dipping to the south and the beds on the southern side dipping to the north. As a consequence the river is eroding approximately along the east-west synclinal axis. Only the upper Tertiary beds are involved in the folding.

To the east and south the axis dies out. It appears roughly to coincide with Wilson's (1963) axis and is thought to intersect the ancestral Waipara River valley, i.e. west of the Teviotdale River (Plate 1).

3.4.2 CASS SYNCLINE

This syncline was first recognised by Speight (1912, p.227) and it is shown on Wilson's map (1963).

The Cass Syncline is distinctly asymmetrical. Beds on the northern limb dip from 26^{o}-45^{o}SE; those on the southern limb dip steeply and are in places overturned. The southern limb, where it has been recognised, is of restricted extent. This is due to the almost complete faulting-out of this limb by the Hamilton Fault. In this southern limb region the syncline has undergone more intense folding than elsewhere, though it still retains its asymmetrical form especially where the Cass Road cuts through the axis, and further to the north near the plunge of the Cass Anticline (Plate 1). In this latter area exposures are covered with thick vegetation which obscures the end of the syncline.

The position of the fold axis is well defined where the Cass Road cuts through the axis, while further to the north the fold axis seems to strike parallel with the trace of the Hamilton Fault, Wilson's map, on the contrary, shows it cutting the fault and continuing on the far side.

3.4.3 BOUNDARY SYNCLINE

Whereas a single syncline is shown associated with the Boundary anticline by Wilson (1963), Carr (1970) maps a pair of synclines relating to his two anticlinal folds on the evidence previously discussed. In the present study only one synline was recognised which separates the Boundary Anticline to the east from the Montserrat Anticcline to the west (Plate 1).

3.4.4 MONTSERRAT SYNCLINE

This syncline which separates the Montserrat Anticline to the south-east from the Black Anticline to the north-west (Plate 1) was shown but unnamed on Wilson's map (1963). The name is taken from the Montserrat Homestead.

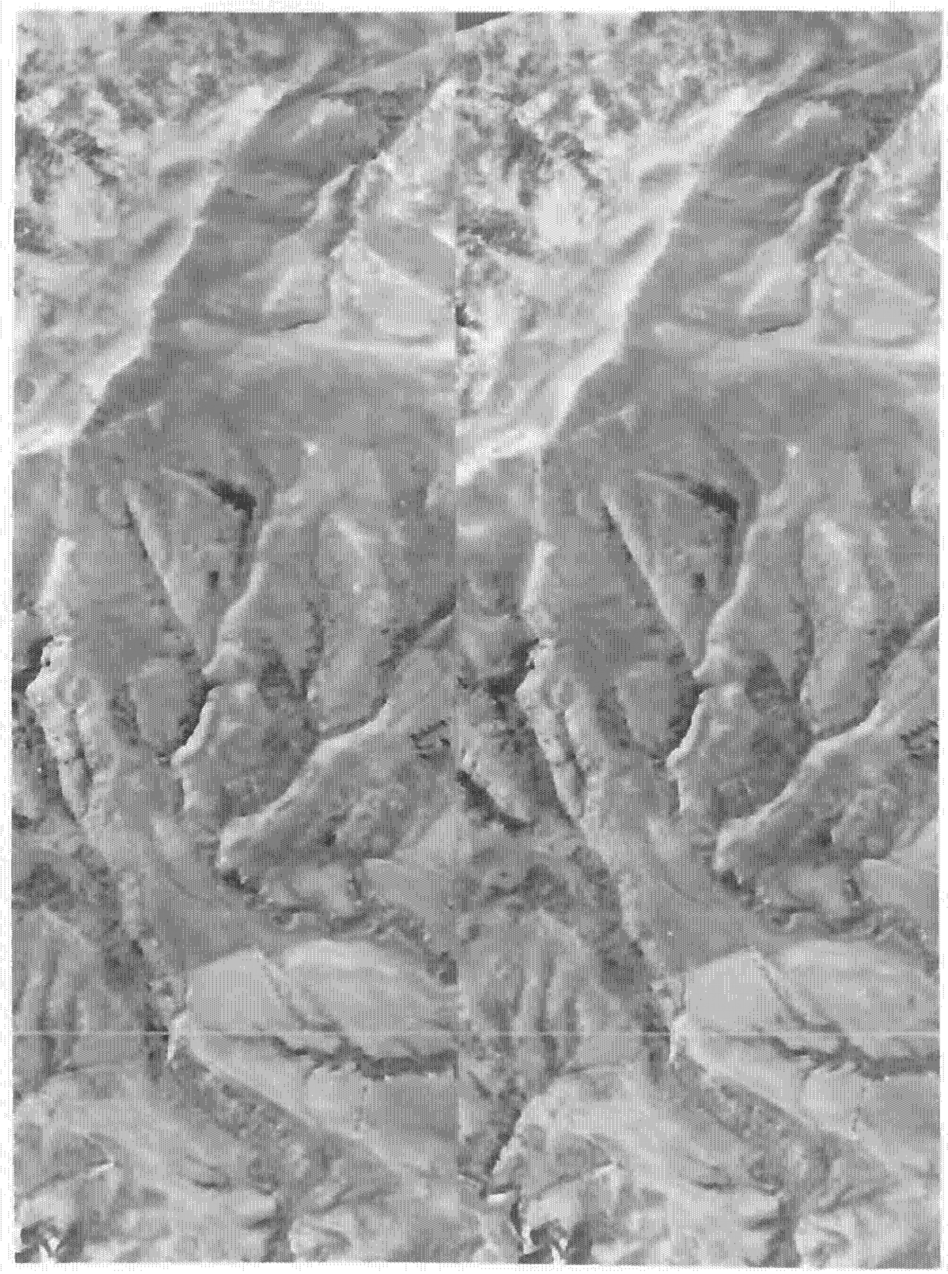
It is easily defined by outcrops on either side of the upper reaches of the Motunau River near the Cotswolds Homestead. The beds on the eastern side dip to the west and the beds on the western side dip to the east. The southeastern limb, where it has been recognised, is of restricted extent due to the almost complete faulting out of this limb further to the north. In large part this structure is represented by the northwestern limb.

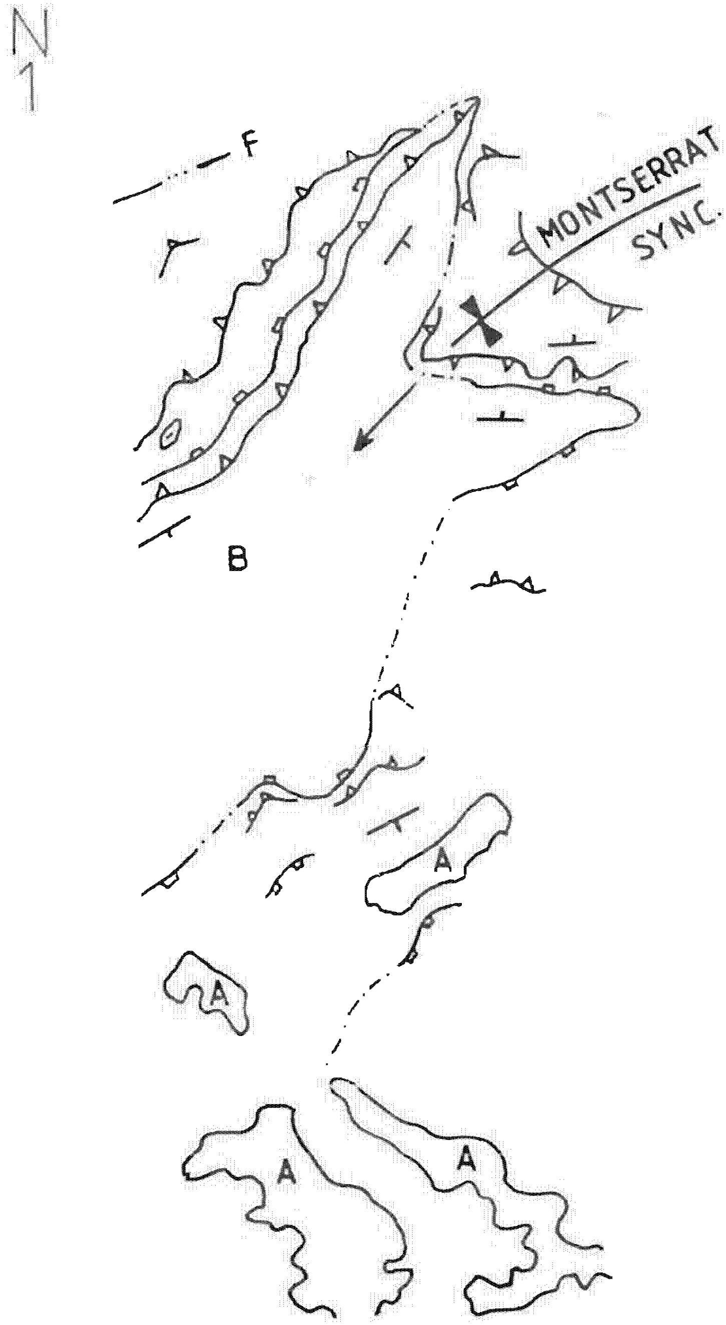
The position of the fold axis is well defined near the Cotswolds Homestead where it runs northeast-southwest and seems to correspond closely to the trace of the north splinter of the Glendhu Fault. The fold is asymmetrical with the limited south-eastern limb having dips that are generally higher than those of the northwestern limb.

At the extreme south end of the Montserrat settlement a well developed syncline forms a further offshoot of the Montserrat Syncline (Plate 1). The result of my survey suggests that this latter syncline is wrongly interpreted on Wilson's map. Observations along the Fitzsimmons Creek and the surrounding area (See section 3.5.6) suggest that a considerable fault exists along this Creek and that it probably extends further to the southeast than previously shown (Wilson 1963). This extension offsets the southern part of the Montserrat Syncline.

Stereomodel 1

Stereomodel depicts the zone around the plunging Montserrat Syncline. The many dip slopes and outcrops in this area, permit the rapid mapping of the essential structural relations. The hogback (area B) of the outer flank of the fold is formed of Omihi Formation (Weka Pass STone). Note the raised, dissected, marine terraces (area A).




N
F
MONTSERRAT
SYNC.
B
A
A
A
A

Additionlly, the position of the axial trace and the direction of the plunge are as shown by Wilson incorrectly interpreted. Although exposures through this area are very infrequent and the extension of the Hamilton Fault disturbs the outcrop pattern in this critical area, a remarkable outcrop exists near the swing of the south-eastern limb of the Cass Anticline, defining precisely the location of the Montserrat Syncline plunge (stereomodel 1).

3.4.5 WAIPARA SYNCLINE

This syncline was first mentioned by Smart (1954). Wilson (1963) suggested that its position is defined by the dipping strata of the Doctors Anticline to the west and the Black Anticline to the east. Wilson (1963) concluded that the deepest part of the syncline was near the junction of the Omihi River and the Waipara River. The Waipara Syncline is distinctly asymmetrical; beds on the north-west limb dip 14^{o}SE; those on the south-east dip steeply and are in places overturned.

No evidence was observed to indicate accurately the position of the axial trace as the middle of the Waipara Syncline is covered by thick late Pleistocene gravels.

3.5 FAULTS

The major and minor fault traces shown on this map (Plate 1) were initially based upon aerial photo interpretation. All mapped fault traces were then walked along their entire known length and supplementary measurements and checks were made. Vertical displacements were measured where possible and photographs were taken.

3.5.1 HAMILTON FAULT

This fault continues in an east-north east direction along the north west limb of the Cass Anticline, except where its southern end swings around the Mount Cass Road to strike roughly south (Wilson 1963).

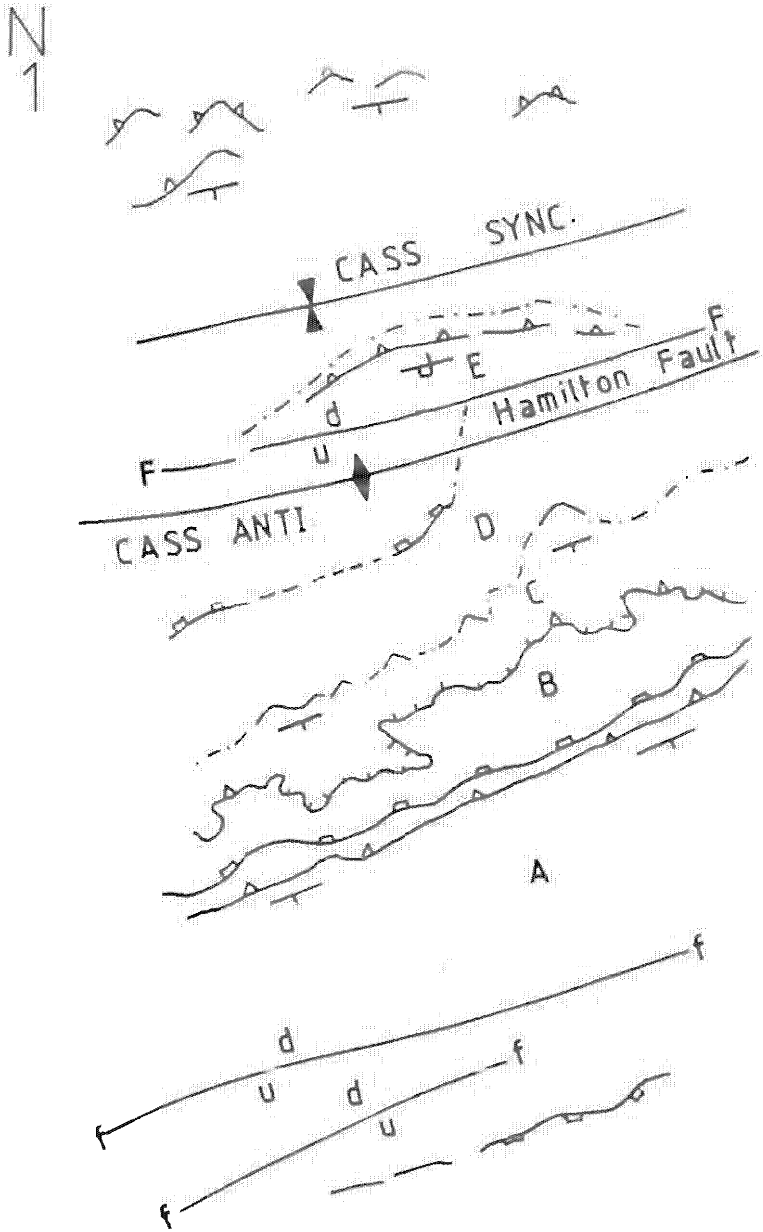
The position of the Hamilton Fault along the northern side of the basement exposure can be placed with confidence from both aerial photographs and field observations. It is visible where breaks in the slope are present, and where fault line scarps are accompanied by a zone of shattering. The fault scarp is well exposed in Smothering Gully (stereomodel 2). The fault separates overturned Amuri Limestone and Weka Pass Stone to the north from Conway Siltstone to the south. Elsewhere the line of the fault is partly concealed by slumping, but the dip is believed to be similar to that of the adjacent Tertiary beds which, dipping south east at between $60^{o}$and 80^{o}, are overturned in several places. The fault is thus reversed and of high angle.

Satellite imagery shows a marked, north east-trending lineation consistent with the continuation of this fault at the eastern extremities of the prominent escarpment of Weka Pass Stone (S68/260148) forming the south-east limb of the Cass Anticline. This lineament is the furthmost extension of the Hamilton Fault, which probably continues under the Waikari Formation west of the Montserrat Homestead to link further northeast with a splinter of the Glendhu Fault. The surface trace of the fault is difficult to determine in this critical area, there being no detectable fault scarps or fault line scarps present.

Near the exposure of Torlesse basement, a fault splinter meets the main Hamilton Fault and runs subparallel to it for 2 km towards

Stereomodel 2

Stereomodel of major and minor faults in differentially eroded moderately dipping strata. Major fault trace (Hamilton Fault) is shown by linear scarps and breaks in slope. The minor faults trending obliquely to strike are revealed in area A. Differential erosion of the exposed strata (A,B,C,D,E) as a key to bedrock identification.


N
CASS SYNC.
F
Hamilton Fault
E
d
u
F
CASS ANTI
D
C
B
A
f
d
u
d
f
u
f
f

the Mount Cass Road (Plate 1). Between these two faults is a remnant of the overturned part of the faulted north-west limb of the Cass Anticline. Maximum vertical displacement of at least 1200 m, occurs north of the basement exposure, where the basement and the Greenwood Formation are in fault contact (Wilson 1963).

The fault cuts the Cass anticlinal axis near its southern end to approximately follow the Cass Road, and rapidly dies out to the south just near the swing of the prominent escarpment of Weka Pass Stone (Wilson 1963). As noted the further east the surface trace of the fault was difficult to determine. Soil creep, rock fall and erosion are likely to have obliterated such features quite quickly.

A. ADDITIONAL MINOR FAULTS

Five distinct reverse faults involved with the south east limb of the Cass Anticline have little effect on the dominant structural pattern (Plate 1). On the basis of fault scarp and stream offset recognition the position of these faults can be placed with confidence both from aerial photographs and field observation. Figure 3.4 shows streams cascading down a fault scarp and quickly cutting steep valleys that erode progressively headward from the scarp into the uplifted block. The resulting valleys are narrow at the lower end and widen upstream into the upthrown block. Dissection by many gullies or valleys causes a scarp to be segmented into a series of triangular facets whose bases are aligned with, or lie parallel to, the fault trace. Two new fault lines were recognized using a gradient index analysis (see Chapter 5). Thus five subparallel faults, involving the southeast limb of the Cass Anticline, form an en echelon system that trends southwest. It is possible that a sixth fault exists running essentially parallel to the swing of the escarpment of Weka Pass Stone

Figure 3.4

Maturely-dissected fault scarps (arrows) forming the south east limb of the Cass Anticline.

Figure 3.5

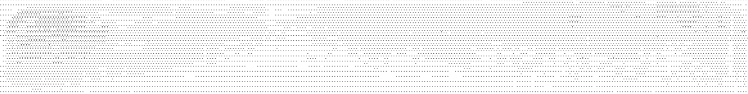
Repetition of Double Corner shellbeds (arrows) and basal Greenwood Formation conglomerate by the reverse fault, at the "Horse Shoe" south bank of Waipara River (S68/133073). View looking south.

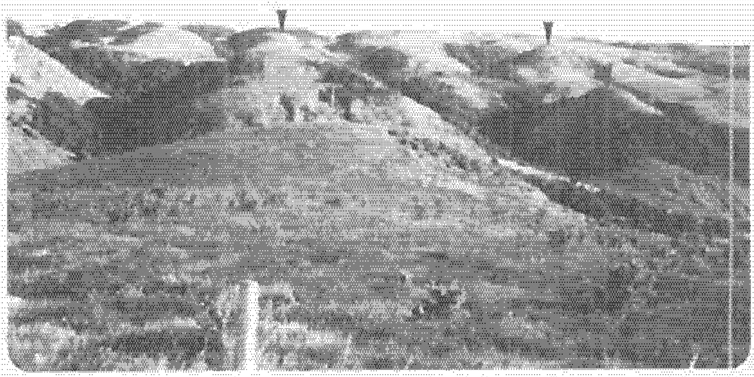
Figure 3.6

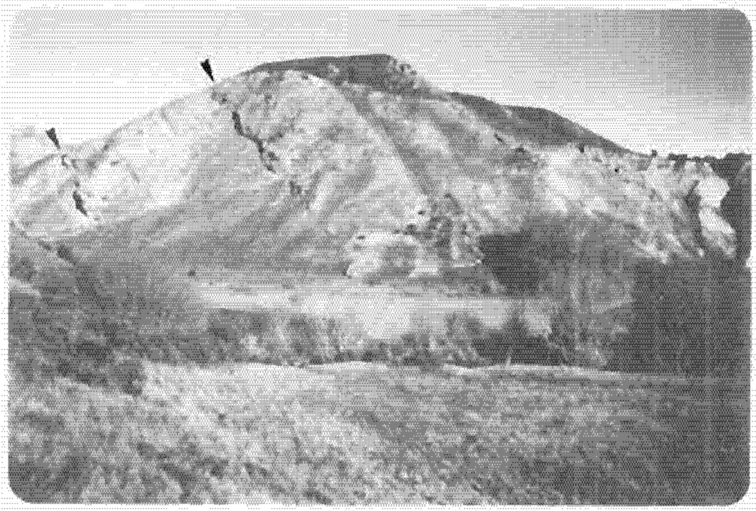
Strike slip fault in the west bank of Waipara River (S68/130054). Note the drag affecting the sandy units (Greenwood Formation) on the upthrow side.

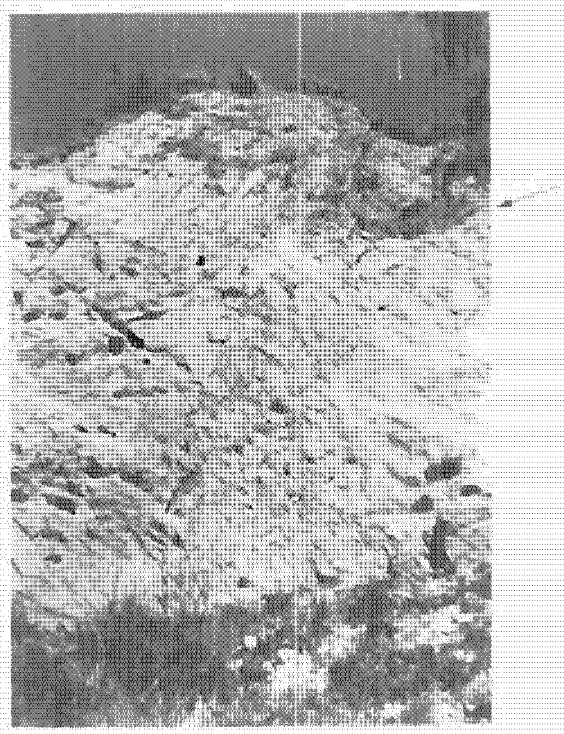
Figure 3.7

Repetition of Waikari Formation sandbeds by the reverse fault (arrow) northeast of Cotswolds Homestead (S68/314170). Note in the lower Motunau River terraces (arrow) there is no trace of this fault.









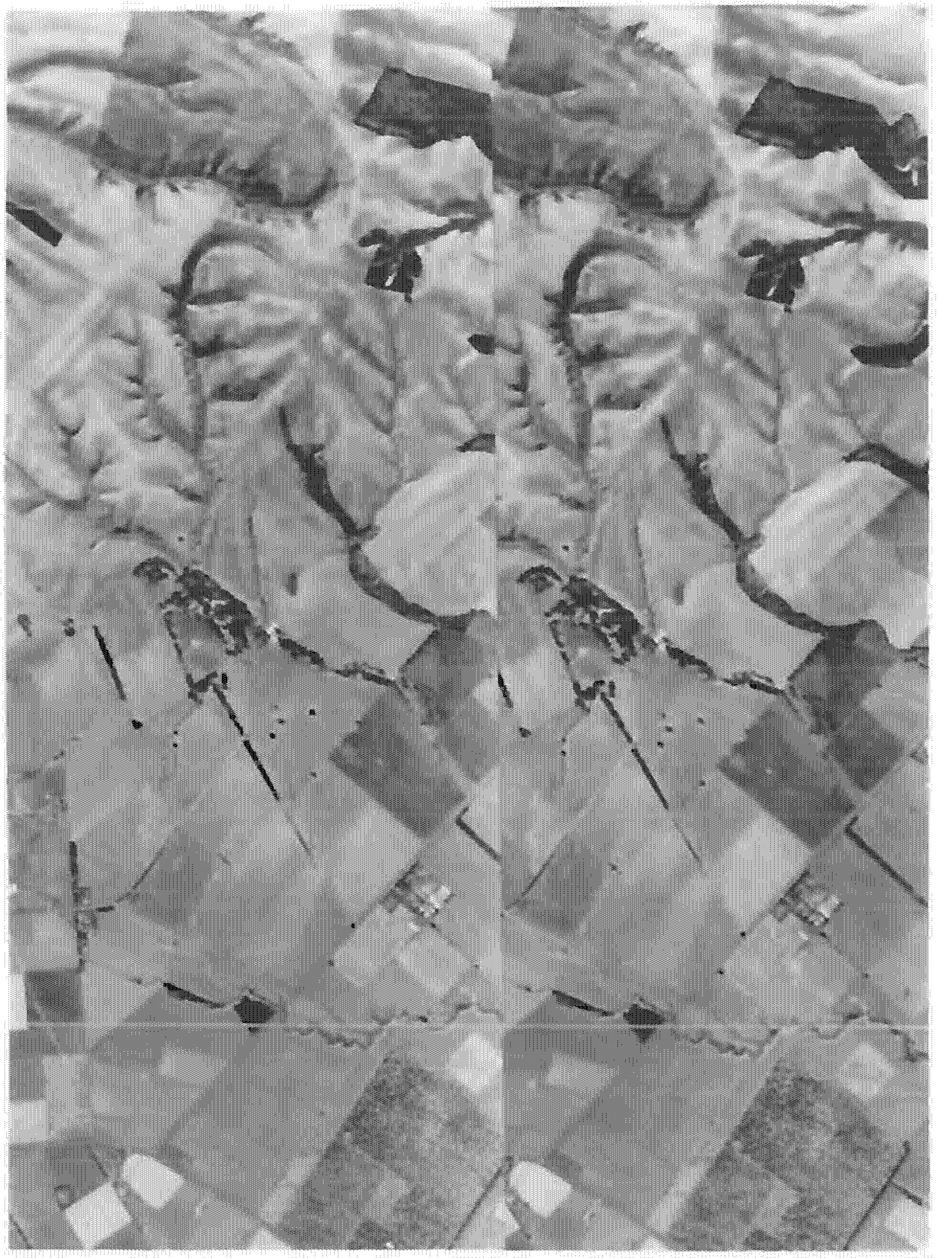
and running toward, but not joining with, the Hamilton Fault. This fault was erroneously interpreted by Wilson (1963) as a highly unusual arcuate continuation of the Hamilton Fault. A more plausible explanation, consistent with all additonal recent data suggests that the main Hamilton Fault died out at the Mount Cass Road (S68/152102), the displacement being picked up by the six subparallel faults arranged en echelon (Plate 1). The parallelism of these faults to each other and their dextral en echelon arrangement suggest a genetic relationship to regional structure.

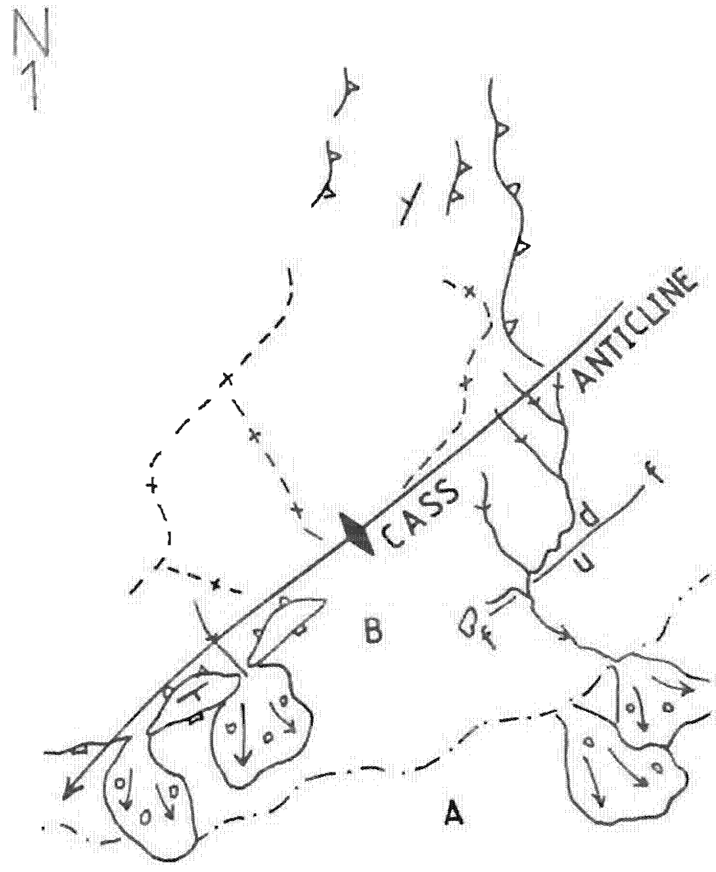
Two faults are exposed where the Waipara River cuts a gorge through the southeast end of the Cass Anticline, north of Greenwood's Bridge. This area was mapped in detail by Gregg (1950), Wilson (1963) and Bradshaw & Newman (1979). The first fault, a strike-slip fault, cuts across the river about 800 m upstream from the bridge (S68/130054). The strata adjacent to the fault on both banks of the river are very disrupted (Fig. 3.5). Anomalies in the terrace profiles across this fault are indicative of late Pleistocene activity (Campbell & Yousif 1985). The fault could not be traced for any distance westward beyond the exposures in the right bank of the river, however, the fault exposed on the east side of the river and striking through the saddle into the Teviotdale Valley duplicates the strike ridge of the Greenwood Formation (Plate 1).

The second fault, which is a reverse fault, runs in a north-east direction subparallel to the Waipara River and duplicates the Double Corner Shell beds. It is clearly visible where it cuts across the incised meander loop of the Waipara River along the north west limb of the Cass Anticline. The repetition of succession is most clearly seen on the right bank of the Waipara River at "Horse Shoe Bend" (Fig.

Stereomodel 3

Typical examples of offset and diverted streams showing right lateral displacement. Note the clearly defined fault line, the direction of offset, lakes, and fault scarp.




N
ANTICLINE
CASS
d
u
f
B
f
A

3.6).

Air photographs reveal evidence of several instances of recent faulting. A step, facing northeast, on the Conway Siltstone Formation east of the Hamilton Fault at grid reference (S68/212130), can be traced north eastwards for about 500 metres. At the foot of this fault trace, sag ponds have developed, as shown in the stereomodel 2.

At the southeast limb of the Cass Anticline, a new fault has been recognized at (S68/115035) using both gradient index analysis and aerial photographs. From stream offset a right-lateral displacement is indicated (Stereomodel 3). Lateral displacement has offset the south-flowing stream, which now follows the fault trace for 250m. A 3 m scarp extends for 200 m along the floor of the valley. To the west a good example of diverted stream. The south-flowing stream is locally entrenched on the upthrow side of the scarp and has cut down in the Kowai Gravels to about 1.5 m below the original valley profile.

3.5.2 OMIHI FAULT

Jobberns (1937b) suggested that a fault scarp existed on the eastern side of the Omihi Valley and that its presence explains the steep face of the coastal hills standing out boldly above the Omihi Valley lowland. He mentioned that no actual fault plane had been located. Jobberns suggested that erosion has taken place such that the steep scarp has retreated a long way from the plane of the ancient fault. Collins (1939) considered the Omihi Fault to be the major tectonic feature of the area but again mentioned that no direct evidence of the fracture was observable. Wilson (1963) also mentioned that the Omihi Fault plane had not been seen but inferred its presence from the disappearance of ridges of steeply dipping Greenwood

Formation and Kowai Gravel conglomerates and also from a generalisation of the regional structure, i.e. that similar steeply dipping northwest limbs of anticlines in the Waipara district are faulted.

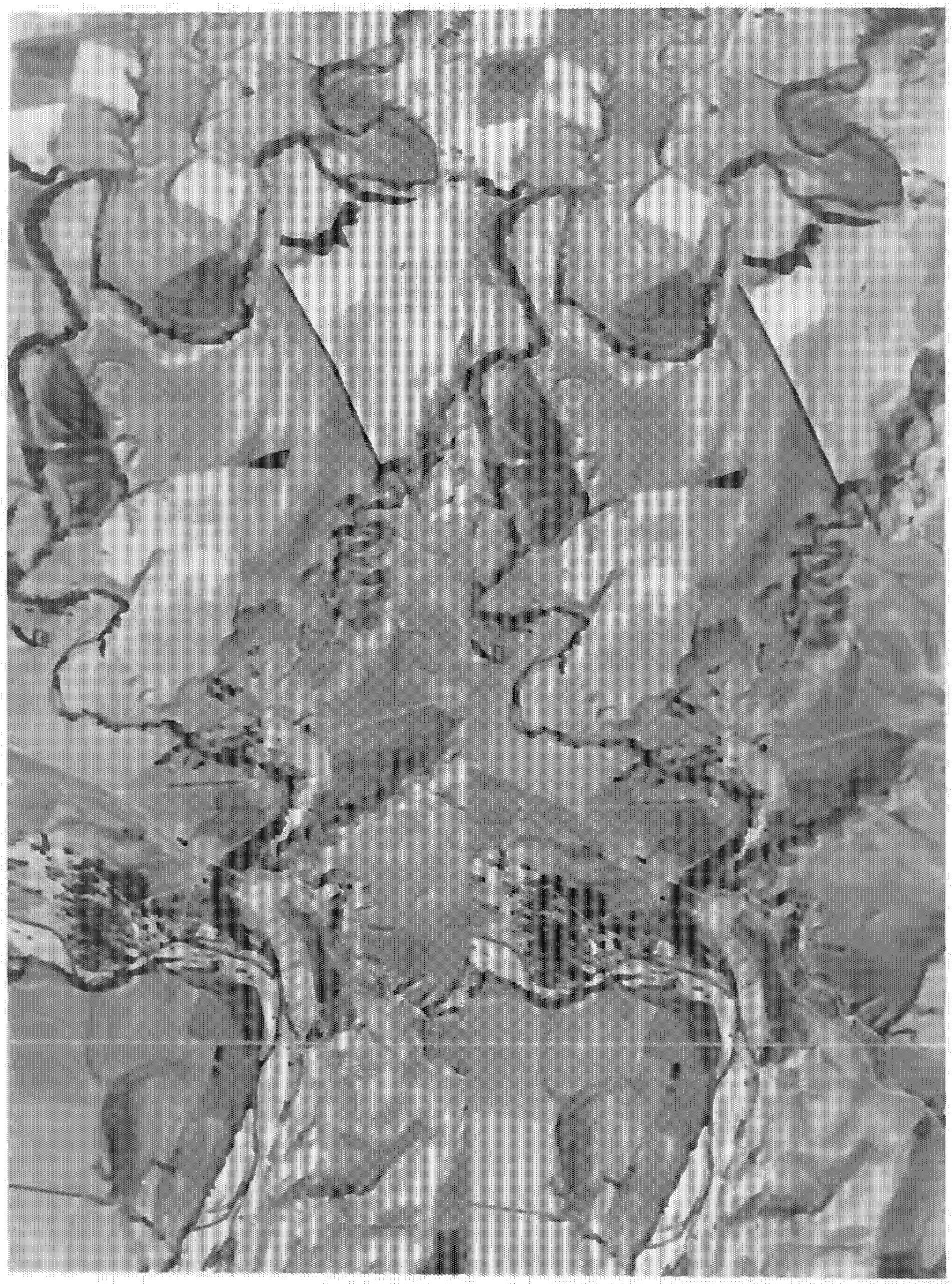
Field observations in this study together with evidence from aerial photographs confirm the existence of the Omihi Fault, but lead to the following re-interpretation.

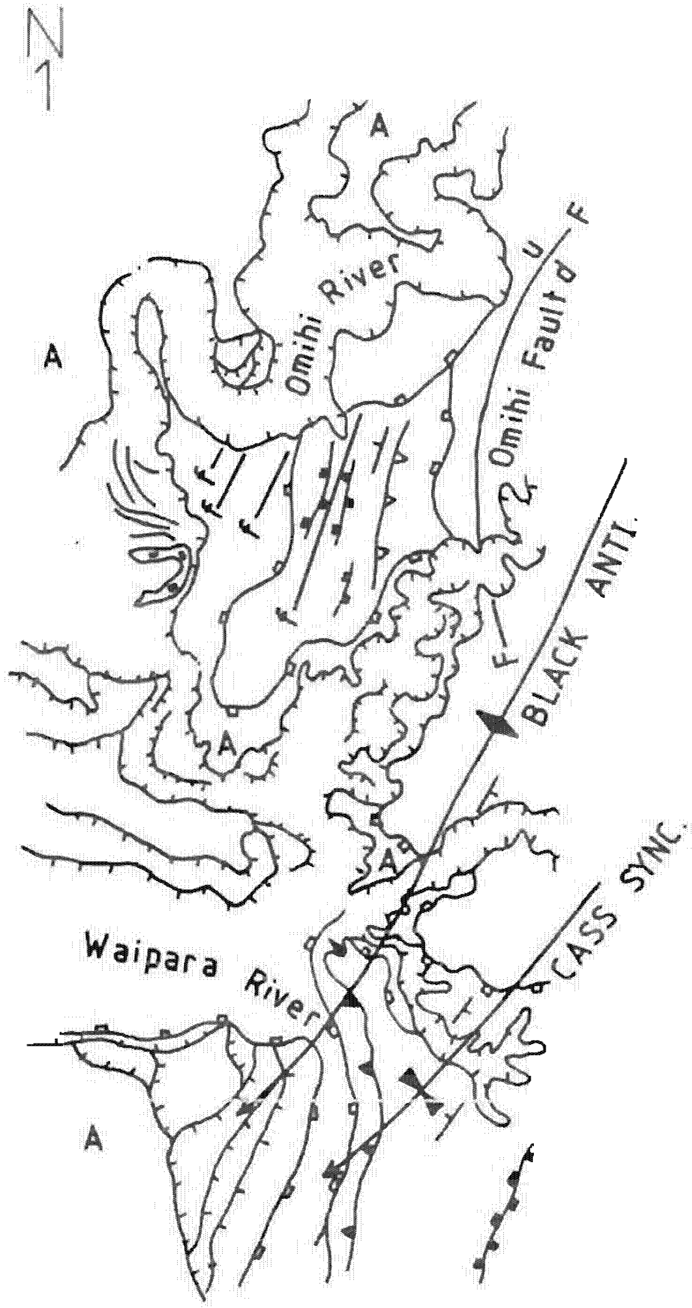
Two faults traces are recognized near the plunges of the southern dome of the Black Anticline (Plate 1). The first fault trace cuts across Limestone Creek about 1250 metres upstream from its junction with the Omihi River (S68/148128). Aerial photographs show a remarkable ridge of a resistant conglomeratic unit of the Greenwood Formation which can be traced for about 800 m to the north from the river bank. The basal ridge appears twice due to fault displacement (Fig. 3.8a). This repetition is most clearly seen on the left bank of Limestone Creek while on the right bank the fault plane is well exposed. Aerial photographs also show a distinct change in the river incision to the south of the fault, in addition to a conspicuous pond west of the conglomerate bed of the Greenwood Formation. These two features can be attributed to the effect of the fault (stereomodel 4). The fault plane strikes at 008^{0} and dips at about 70^{0} to the west and is thus reversed and of high angle. It could not be traced beyond the exposures in the banks of the river due to the thick slope-wash deposits.

The second fault is inferred to run along the north western end of the southern dome of the Black Anticline near to the swing of the fold axis (S68/186164). Although the fault plane is not exposed, aerial photographs and field evidence define marked local variations

Stereomodel 4

Stereomodel of drainage anomalies in a low to moderate relief terrain. Streams incision and drainage characteristics are the result of late Quaternary active folding and faulting. Note the high sinuosity, ox-bow lakes, flight terraces, and progressive westward shifting of the stream course related to structural uplift to the east.





N
A
A
Omihi River
Omihi Fault
u
d
F
f
f
f
F
BLACK ANTI.
A
A
Waipara River
CASS SYNC.
A

in both amount and directions of dip that are explicable by no other means. In addition a window of highly sheared Amuri Limestone Formation outcrops adjacent to the fault (Stereomodel 5).

The Omihi Fault, as currently interpreted (Plate 1), is restricted to the southern dome of the Black Anticline. The extreme limits of the fault are defined by the two faults traces mentioned above. Roughly in line with, but further to the northeast there is another fault bordering the northern dome of the Black Anticline, but this is isolated from the major fault.

From stratigraphical and field evidence overturning has been observed to occur widely in the north-western limb of the southern dome of the Black Anticline. This is well illustrated in Plate 1. This overturneis considered to be of a subordinate nature, being dominantly controlled by faulting during deformation.

A. ADDITIONAL MINOR FAULTS

Observations of the east bank of the Omihi River (S68/138134) suggest that faulting is now the dominant mode of deformation occurring in this part of the northwest limb of the Black Anticline (Fig. 3.8b). Fault displacement occurs in the Kowai Gravels and in some cases displacement of the overlying Canterbury Gravels was seen. Fault displacement, therefore, postdates the last episode of folding. The most prominent fault plane exposed at the river bank is subparallel to the bedding in the Kowai Gravels and has a dip of 60^{o} W, striking 012^{o}. Throw is estimated to be 10 m based on the displacement of the strath surface. In addition, several smaller faults with displacements in the order of 50 cm also displace the strath surface (Fig. 3.8 d&e).

Figure 3.8a

Repetition of conglomeratic units (arrows) by high angle reverse fault (Omihi Fault). view looking ENE. Note the lake at the foreground formed by the fault movement.

Figure 3.8b

Structural cross-section along the northwestern limb of the Black Anticline (southern end) with photo locations (A,C,D,E).

Figure 3.8c

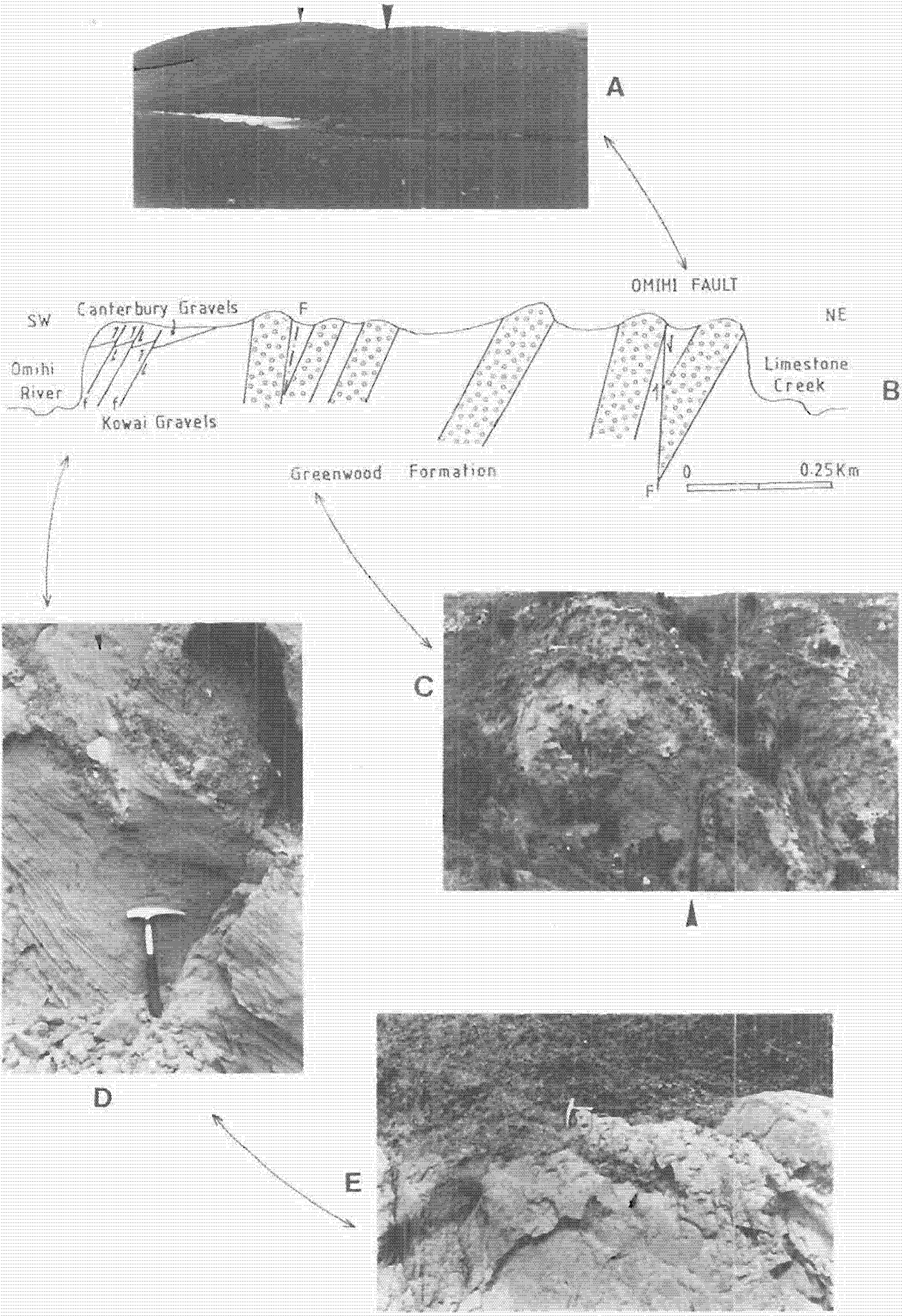
Exposure of inferred fault plane between the underlying steeply dipping Greenwood Formation and the younger slope wash deposits. View looking SSE.

Figure 3.8d

Kowai Gravel units displaced 40 cm by reverse fault (Dip 70° SW). Left bank Omihi River (S68/140134). View looking SSE.

Figure 3.8e

Small thrust fault (Dip 40° SW) displacing Canterbury Gravel strath surface by 52 cm (approx.). View looking SSE.


A
OMIHI FAULT
SW
Canterbury Gravels
F
NE
Omihi
River
Kowai Gravels
Limestone
Creek
B
Greenwood Formation
0
0.25Km
F
C
D
E

Three parallel ridges were recognized between the new position of the Omihi Fault and these recent faults (Fig. 3.8b). The lower ridge intersects a southeast trending gully eroding back into the low hills to the east of the Omihi River. In the gully, shellbeds of the upper Greenwood Formation outcrop as vertical to slightly overturned deposits (Fig. 3.8c). Although a fault plane was not exposed, a fault is suspected between the uppermost shellbed and the overlying gravels, an inferrence based on the abrupt change in the dip of the strata as well as on the change in dip direction across the fault. This suggests that faulting is due to compressive stress causing displacements along suitably oriented planes of weakness, in this case bedding planes.

Obvious reverse and normal faults have offset recent river deposits on the east bank of the Carrington Creek (S68/142110) parallel to that of the Omihi River. This will be discussed in detail in Chapter Six.

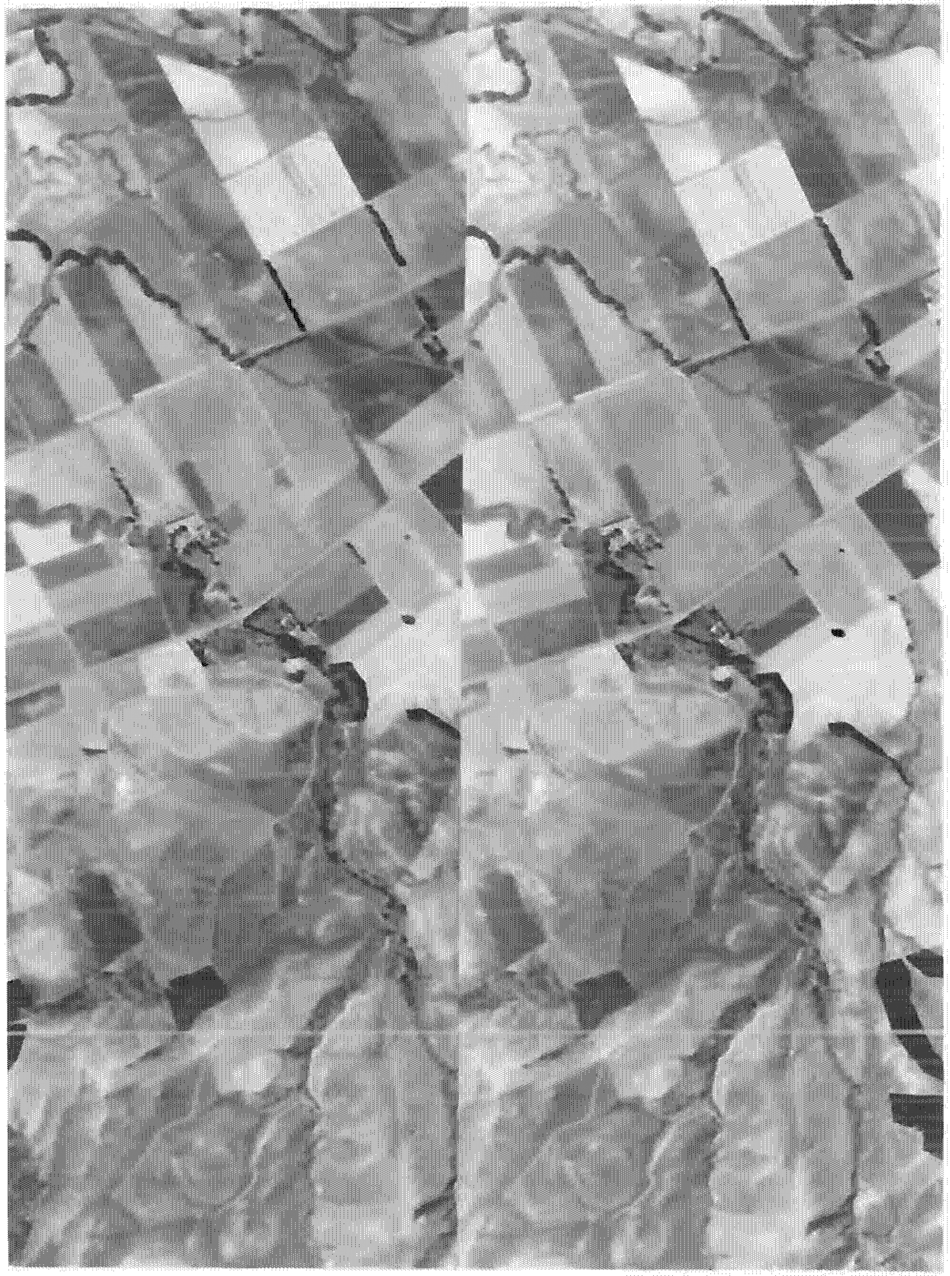
3.5.3 GLENDHU FAULT

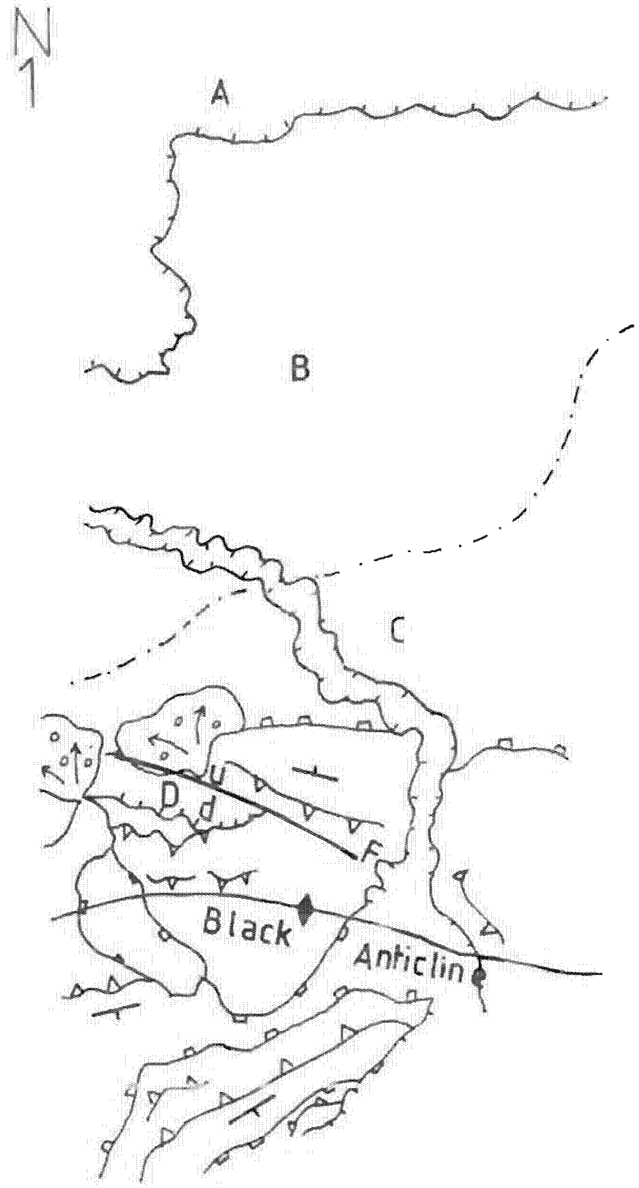
This fault was named on Wilson's map. It is subparallel to the north western limb of the Montserrat Anticline.

The position of the ends of the Glendhu Fault can be placed with confidence both from aerial photographs and field observations, at points where breaks in slope are present, together with fault line scarps. At its northern end, there is a remarkable break in slope where basement and Greenwood Formations are in fault contact. The maximum vertical displacement is at least 1200 m (Wilson 1963). At its southern end, the fault line scarp follows a marked, north-east trending lineament which dies out further to the south. Beds are

Stereomodel 5

Stereomodel of faulted anticline in low to moderate relief terrain. Faulting recongnised by changes in dip slopes, and exposure of older formation through younger cover at D. Linear tonal contrasts suggest a north west-trending fault. This area shows the transition zones between Omihi Valley (A), Canterbury Surface (B), Foot slope (C) of the surrounding lowlands.




N
A
B
C
D
u
d
F
Black
Anticlin

transected by this lineament. This is well illustrated in stereomodel 6.

The surface location of the fault is difficult to determine between these two end points, there being no detectable fault scarps or fault line scarps present. Wilson's map shows the presence of two fault splinters which meet the main Glendhu Fault and run subparallel to it toward the Montserrat Homestead for about 4 km. The first splinter is seen to coincide closely with the Motunau River which continues in an east-north east direction along the north west limb of the Montserrat Syncline. The fault plane is well exposed about 1 km northeast of the Cotswolds Homestead (S68/314170). The fault plane strikes at 240^{o} and dips at about 48o to the east (Fig. 3.7). The fault is thus reversed and of high angle. This fault may continue to the southwest, close to the Hamilton Fault, and if continuous would form a major fault trace in the area. The second splinter is shown running subparallel to the first one, but the surface trace of this fault has not been preserved. However the dip is believed to be similar to that of adjacent Tertiary beds which, overturned at its southern end, dip to the northwest. Both these splinters die out near the Montserrat Road.

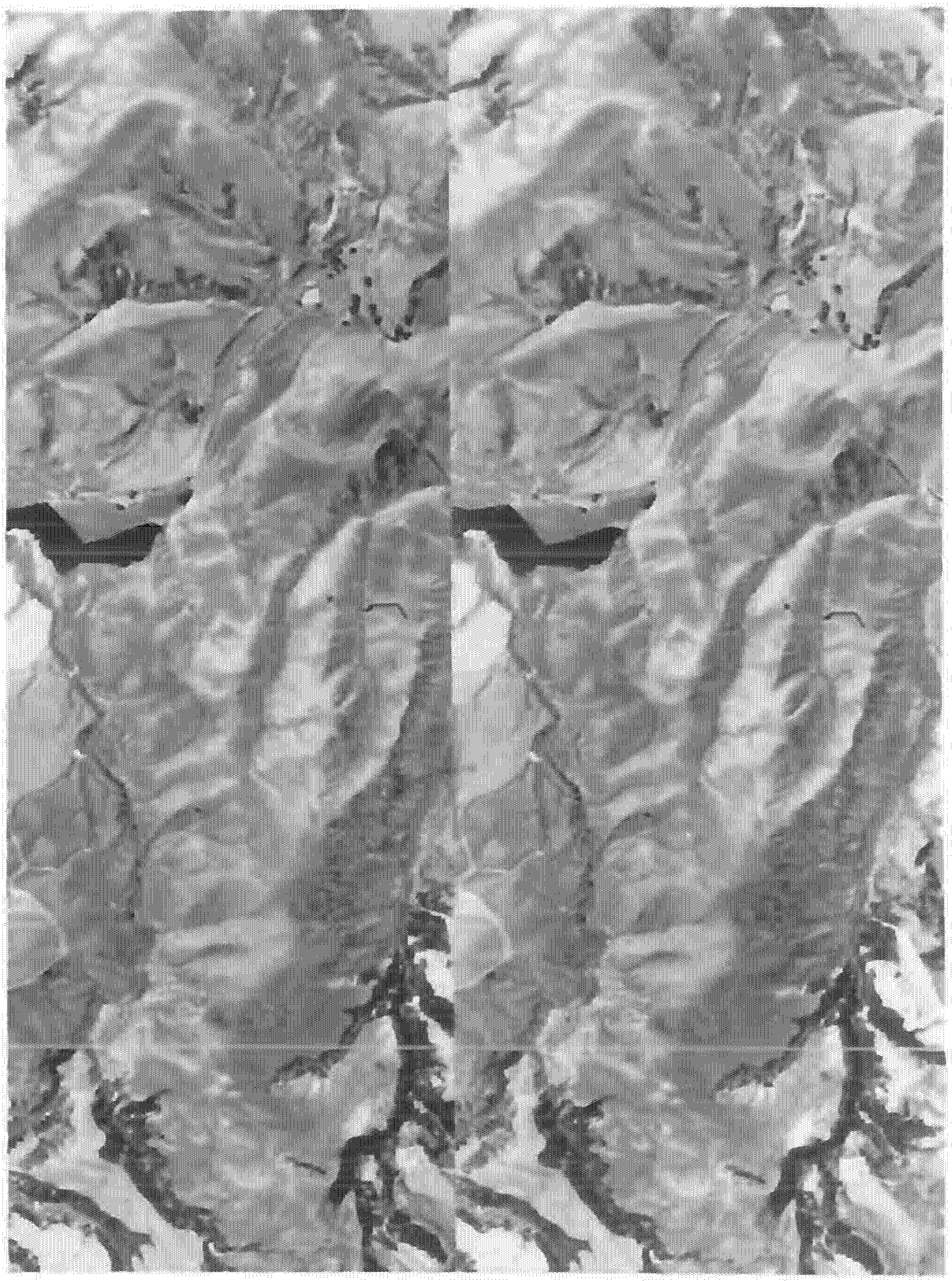
3.5.4 LIMESTONE CREEK FAULT

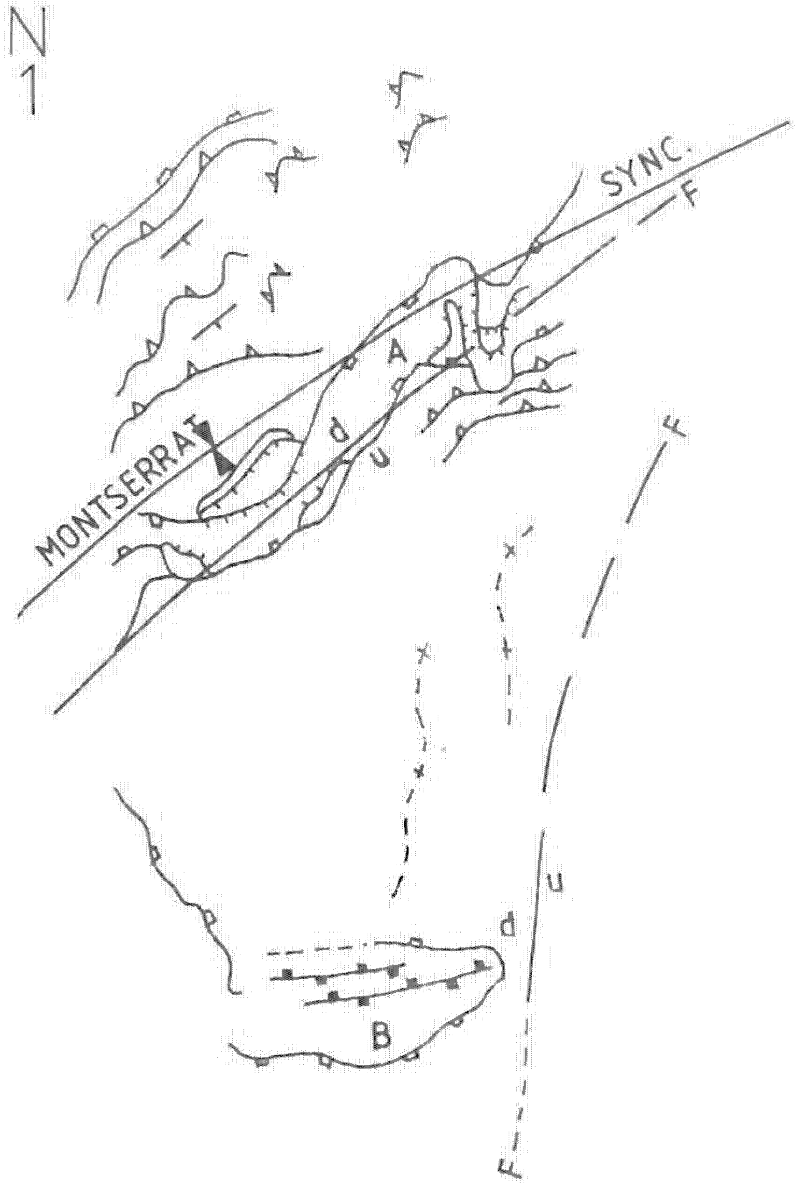
This fault was shown but unnamed on Wilson's map. The name is taken from Limestone Creek. The fault affects the limestone of the Weka Pass Stone Formation associated with the southern dome of the south eastern limb of the Black Anticline.

Aerial photographs show a distinct break in slope between the limestone and the adjacent lithologies due to the vertical

Stereomodel 6

Stereomodel of a maturely dissected faulted syncline. Two splinters of the Glendhu Fault cross the dipping beds in this area. Note the pronounced intersection of the strike like ridges (area B) with the drainage lineament, suggesting a fault trace.




N
SYNC.
F
A
d
u
MONTSERRAT
F
u
d
B
F

displacement by the fault. The fault trace clearly follows the creek direction but is obscured by overturning and slumping.

3.5.5 MONTSERRAT FAULT

This fault was shown but unnamed on Wilson's map. The name is taken from Montserrat trig. It occurs along the southeast limb of the Montserrat Anticline west of the Montserrat Landslide where a fault scarp is present accompanied by offsetting streams. It swings towards the north, and dies out before crossing the fold axis. There is a real possibility however that this feature is in fact a slump scarp. This expla nation could be accounted for by internal slumping of the Weka Pass Stone Formation. In the steep coastal hills near Montserrat Homestead, the spectacular hogbacks of the Weka Pass Stone and Amuri Limestone Formations, associated by gravity, have produced a similar intensively slumped topography (Stereomodel 7). The soft, easily eroded and frequently slumped mudstones are the dominant slope-forming materials in this area (see Chapter 4). This slump affects the Weka Pass Stone outcrop and has nearly blocked many of the small streams that now follow the main slump scarp. However detailed mapping is needed to solve this problem.

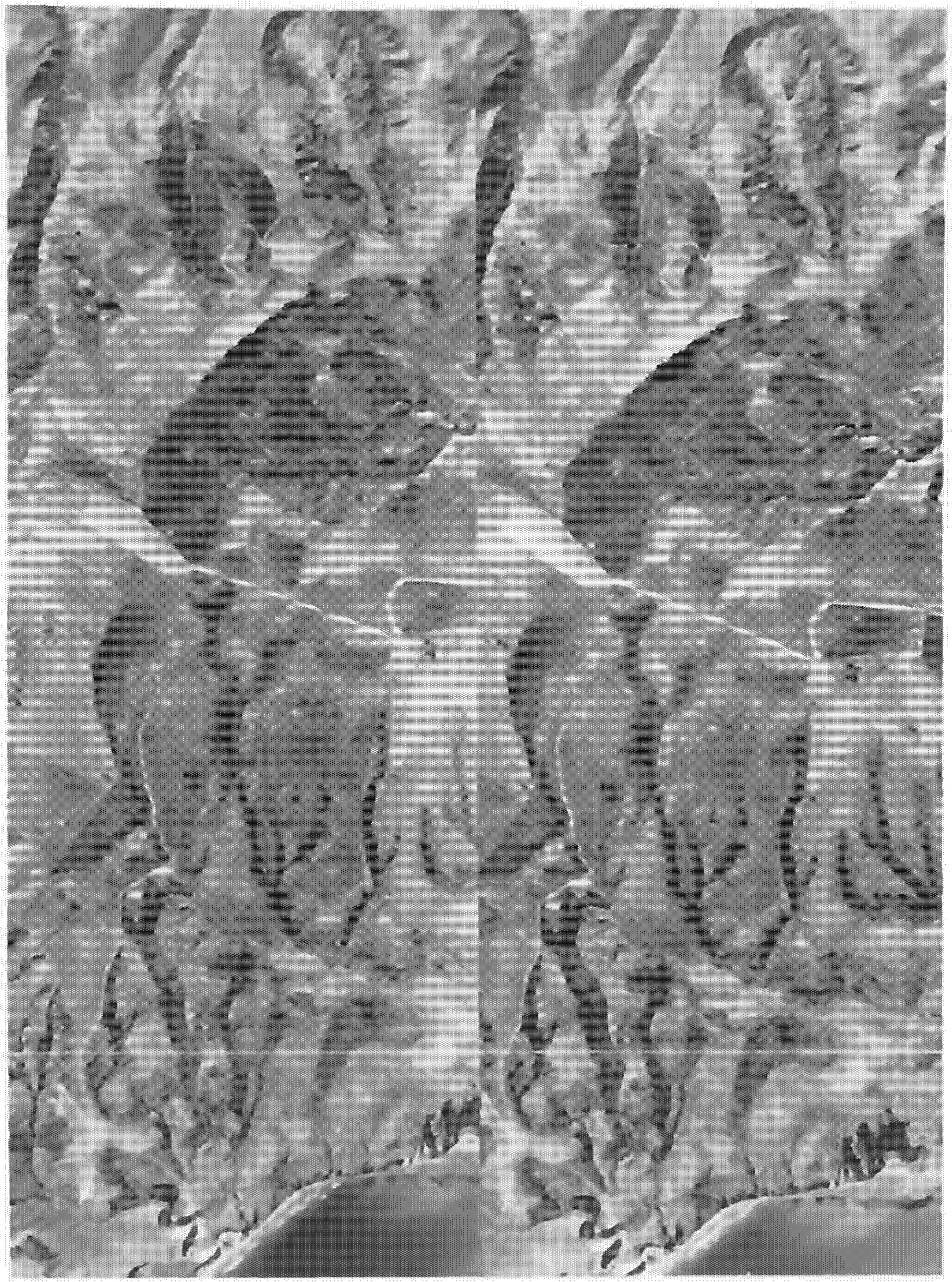
3.5.6 FITZSIMMONS FAULT

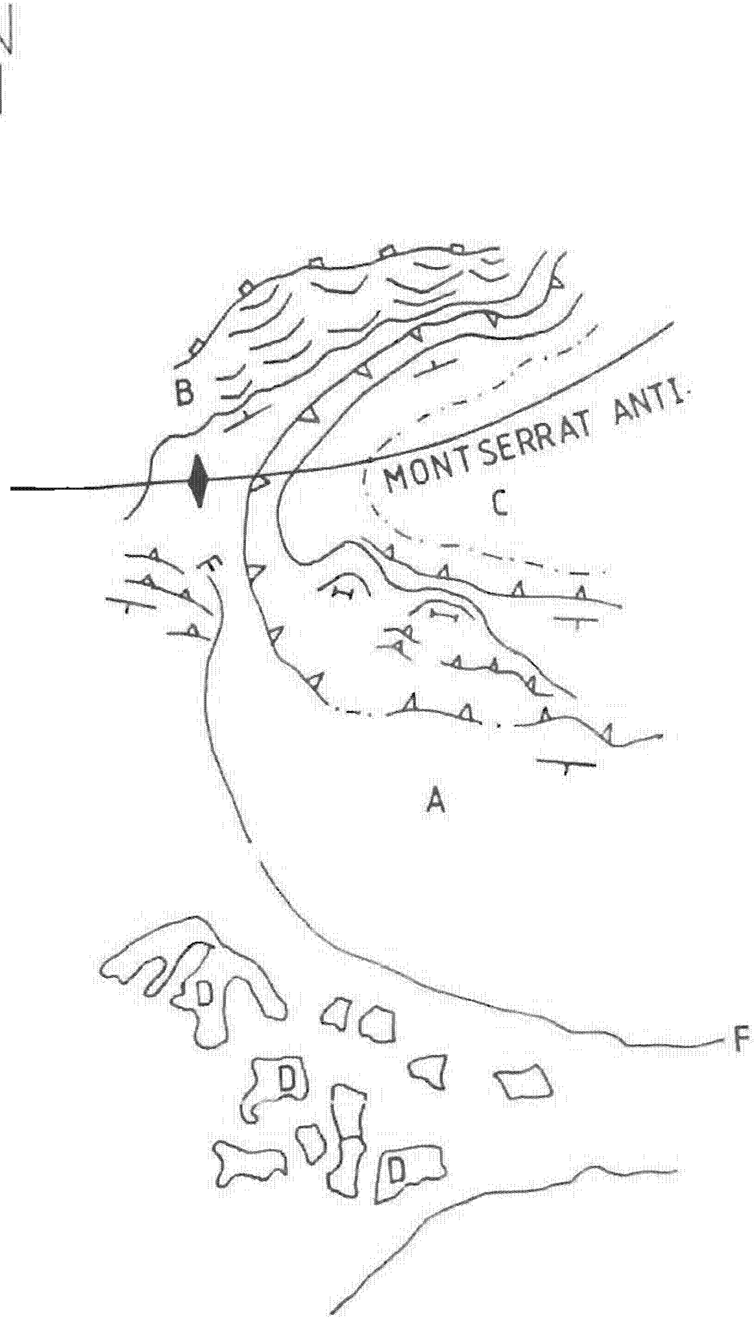
This fault was inferred from Wilson's map. The name is taken from the Fitzsimmons Creek. It cuts across the north eastern limb of northern dome of the Black Anticline and it swings towards the northeast where it dies out before crossing the fold axis. It also dies out towards the south (Plate 1).

This fault is a dominant structural element in the area. It has

Stereomodel 7

Stereomodel of the differentialy eroded nose of a plunging asymmetric anticline. Lithologic differences between formations (A,B & C) are clearly evident by contrasts in photographic tone and erosional characteristics of terrain. Note off-set streams, as at area A, marked by fault trace, are probably large scale landslide. Note also the raised, dissected, marine terraces (area D).




N
B
MONTSERRAT ANTI.
C
F
A
D
D
D
F?

brought about a remarkable swing in the bed dip and general outcrop pattern in the vicinity of the fault. The strike and dip varies from 065/19° SE to 292/42° S. A zone of shattering is exposed at grid reference (S68/274158) and further to the west along the valley, local indurated grey siltstone beds are overturned. At this locality the dip of the siltstone is 64° N, striking 120°.

The fault trace follows a marked lineament generally coincident with the Fitzsimmons Creek.

3.5.7 HAPPY VALLEY FAULT

This fault was inferred but unnamed on Wilson's map. The name is taken from the Happy Valley Homestead. It strikes east-northeast along the south east limb of the Montserrat Anticline about 1.5 km north of the Motunau River. It cuts across the south east limb and dies out before crossing the fold axis, while probably linking with Carr's fault further to the north outside the study area.

Over the full length of the fault, formations are noticeably displaced and a fault line scarp is clearly recognisable from aerial photographs and field observation. The fault is oblique-slip, of low angle, and is well illustrated in the stereomodel 14.

CHAPTER FOUR

DESCRIPTIVE GEOMORPHOLOGY OF THE STUDY AREA

4.1 INTRODUCTION

4.1.1 OBJECTIVES AND RECENT DEVELOPMENTS IN TECTONIC GEOMORPHOLOGY

Neotectonic studies incorporate a variety of disciplines such as geophysics and geodesy and one important area is tectonic geomorphology. This branch of the discipline is here defined as the delineation and study of landforms that have been modified, or resulted from, crustal movements and includes the study of geomorphic processes that have been affected or initiated by tectonism. A problem implicit in tectonic geomorphology is the distinction between landforms modifed by tectonism and landforms merely the result of non-tectonic geomorphic processes. The first textbook on the subject has recently been published (Morisawa & Hack 1985), and we are now at a stage when we can use our knowledge of landforms to throw light on problems in tectonics. It is clear that drainage patterns, for example, can be shown to have either pre, or postdated tectonic movements, and in this way the chronology of landform development constrains the timing and nature of these movements. Tectonics can no longer ignore morphological evidence.

Geomorphologists have put considerable effort into the study of the relationships between geology and landforms, and we can now be fairly confident about predicting the effects of lithology and structure on the development of drainage, and of landforms in general (e.g. Cotton 1941 & 1958, Holmes 1944, Thornbury 1969, Easterbrook 1969, Ollier 1981, Soons & Selby 1982, Selby 1985). Recent studies have focused on the relative rates of uplift associated with different landscape assemblages, relative and/or absolute rates of horizontal and vertical displacement along faults, and times of last surface rupture (Wallace 1977a & 1978, Keller et al. 1982a & b, King & Stein 1983, Nash 1984). Analysis of drainage patterns as a means of delineating the evolution of landscapes has not received the same attention, but pertinent recent studies include those of Schumm (1981), Adams (1980b), Ouchi (1985) etc. documenting the response of fluvial processes to gradient changes. If this is indeed the case, then the relationship between such gradient changes and the recent behaviour of river systems should be carefully examined. Through river studies it may be possible to advance further ideas on the neotectonic behaviour of the crust.

Whereas many would perceive neotectonic studies as simplistic recent structural geology, more recent research in geomorphology is in fact beginning to demonstrate that the opposite is true (Doornkamp 1986). Far from being simplistic, the way forward in neotectonic research is very complex. There are at least three reasons for this: 1) new data allow greater precision; 2) geomorphological theory has changed; 3) new approaches are emerging.

Presently available Landsat imagery has been shown to have value in tectonic studies, though for large-scale mapping more conventional data sources such as black and white air photographs will remain important. The present study illustrates a traditional approach to the use of remote sensing through air photographs in an investigation of the influence of neotectonism on geomorphic development within the Waipara region.

4.1.2 CATEGORIES OF GEOMORPHIC DATA PERTINENT TO TECTONIC INTERPRETATION.

The morphologies of landform elements such as hillslopes, longitudinal profiles of channels, mountain fronts, fault scarps, alluvial fans, marine and fluvial surfaces etc provide data which, when quantified in some manner, can be useful in evaluating the effect of base-level change. Bull and McFadden (1977) use quantitative morphologic parameters such as mountain front sinuosity, valley width, basin elongation in conjunction with qualitative parameters like fan morphology, to categorize vertical tectonic activity along mountain fronts bounded by depositional basins.

By using this approach Bull and McFadden provided the following classifcation for the case of range bounding faults:

Class 1. <u>Active tectonism</u>; associated landforms include unentrenched alluvial fans, elongated drainage basins, narrow valley floors, steep hill slopes.

Class 2. <u>Moderate to slightly active tectonism</u>; associated landforms include entrenched alluvial fans, large drainage basins, steep hillslopes, valley floors wider than their floodplains.

Class 3. <u>Tectonically inactive</u>; associated landforms include pedimented mountain fronts and embayments, steep slopes only on resistant strata, a few large and integrated valley systems in the mountains.

This study by Bull and McFadden (1977) established the landform characteristics which may be applicable in other fault-generated mountainous terrain such as those of the Waipara region.

Hack (1973) used the logarithmic downstream increase in stream discharge, and the slope of a given reach, to define a stream-gradient index, which is the product of the slope of a reach times the logarithm of the length from the headwaters divide (see Chapter 5, section 5.4.4). For comparisons of channel segments along a single channel, the gradient index is an easily measured and useful paramter. A joint study of stream gradients by way of the gradient index requires that (GI) values be plotted on a map and contoured if possible or have zones of steepness identified. Abrupt steepening of the contours show steepening of stream channels gradient and if regional in extent, they are seen as linear features on the gradient index map. Local steepening of stream channel gradient may be due to geologic contraints but regional steepening is more likely to be tectonic in origin and this is discussed in Chapter Five.

Marine and fluvial terraces can be used as well-defined key surfaces to detect tectonic deformation in uplifted areas (Bull 1984). Much of the significant research of the last decade has centered about dating of marine terraces and deciphering Quaternary sea-level fluctuations (Machida 1975, Pillans 1983, Kennedy et al. 1982, Bull & Cooper 1986). A basic assumption of the sea-level curves of Bloom et al. (1974) or Chappell (1983) is that the rate of uplift at a transect on the New Guinea coast has been constant with time. Their work provides a basis for estimating rates of uplift, folding, and faulting on other coast lines with marine terraces.

Strath, fill, and cut stream terraces (Leopold and Miller 1957) represent time lines along valleys because they are formed during periods of equilibrium or threshold conditions (Bull 1979). Thus, stream terraces record former longitudinal profiles of valley floors that subsequently may be deformed by either vertical (Keller et al. 1982, Burnett & Schumm 1983, Yeats et al. 1981) or horizontal (Lensen

1968) earth deformation. This study area is ideal for developing a better understanding of the relations between streams and uplift. The drainage network has evolved in close association with active folding. This led to waves of landform modification transmitted along the drainage network, for in essence, warping does more to river gradients that it does in any direct sense to hill slopes. An examination of the down-valley variations in river channel slopes throughout the lower Waipara area may therefore reveal a great deal about the nature of warping, its magnitude, frequency and spatial variability within the area. Further discussion is reserved for Chapter Six.

Rivers respond to active tectonic movement in ways that are dependent on their hydrological characteristics and on the types of deformation. Alluvial channels are sensitive indicators of change in hydrology and sediment load and type, from a variety of causes; the degree to which the tectonic factor can be identified, has not been extensively investigated, particularly with respect to active folding. There are a few notable contributions, such as Tator 1958, Welch 1973, Adams 1980b, Schumm 1981, Russ 1982, King & Stein 1983, Burnett & Schumm 1983, Schumm et al. 1982, and Ouchi 1985.

River sinuosity studies (such as those of Adams [1980b] in the United States) have been used to identify surface tilting by current and recent tectonics. The conceptual link between sinuosity and tilting involves the analysis of the river slope required to carry the sediment load. Since meandering rivers tend to establish and maintain the equilibrium gradient required to carry their sediment load, any change in the gradient of the valley floor (e.g. induced by tilting) can be accommodated by the river through a change in sinuosity. A study of sinuosity changes may provide significant ideas on tectonic deformation, but the Waipara River and its adjacent smaller streams appear to represent a system which has evolved further than Adam's examples and probably reflects more complications in the deformation which are discussed in Chapter Six.

New Zealand rivers in the area to be discussed respond rapidly to changes in tectonic and climatic inputs, and thus are ideal for the development of new concepts and tools in tectonic geomorphology. In this project, seven streams were studied in detail and discussed in Chapter Six. From south to north, they are the Rakaia River, in mid Canterbury, the lower Waipara River and its tributaries Carrington and Yellow Rose Creeks, the Teviotdale River and Kate Creek, all near the lower Waipara gorge area, and the Motunau River to the north of the study area. These streams provide good examples of the nature of channel response to actively growing folds and faults and despite the differences in size and erosive power, they all show features in common. To understand the approach used in this study to problems in tectonic geomorphology, one must also understand some basic concepts in geomorphology. It is desirable that they be explained at the outset.

In a practical sense, tectonic geomorphology seeks to interpret the relation between surface energy in the form of streams, runoff etc and surface material by way of resulting landforms. Basic to interpreting geomorphic data is some knowledge of how energy and landforms are related.

Mapping the distribution and the chronology of the fluvial and marine surfaces together with surveying of tilted strath and aggradation terraces, can provide data in the style and rate of active tectonic processes deforming the geomorphic surfaces.

4.1.3 REVIEW OF PUBLISHED GEOMORPHIC CLASSIFICTION AND CARTOGRAPHIC SCHEMES

Geomorphic mapping is presently considered an essential pre-requisite for applied geomorphological research. This concept was not initially accepted; in fact, the methodology was largely developed from the nineteen-fifties onwards and the importance of the techniques involved for surveys and development are not, even to date, universally understood (Verstappen 1977).

The aim of a geomorphological mapping is to provide concise and systematic information about landforms and related phenomena. Landforms can be classified on the basis of several categories of criteria such as morphologic, geologic, chronologic, genetic, and dynamic (Zuidam & Zuidam 1979). A comprehensive description of landforms may necessitate incorporating all the above five aspects. Outlining major landform units and plotting minor forms will be of little use unless accompanied by a thorough analysis leading to a reconstruction of the geomorphological development, identifying the geologic, climatic and other factors prevailing at present and in the past. Therefore, a comprehensive geomorphological map should combine as many of these factors as possible.

Many countries developed their own systems of geomorphological mapping. A comparison of the various mapping systems and legends has been attempted by some geomorphologists (Gilewska 1967, Van Dorsser & Salome 1973, Salome & van Dorsser 1985). Without an extensive description of these systems a number of similarities and differences have been compiled in two tables (Table 4.1 & 4.2) which permits comparison of the systems. For example, the French system is based upon geological structure, climate independent (marine, fluvial, etc.) and climate controlled processes. Colours are used for geological structure and landforms resulting from geomorphological processes. In this system, only coloured symbols are used. Thus it is possible to mix these symbols to indicate polygenesis. The German system includes all elements of landforms using many symbols and colours, and in particular, it emphasizes both morphology and current and past geomorphological processes. As a result of too abundant information, the map loses clarity and legibility. The legend of the geomorphological map of the Netherlands system is structured in such a way that the relief classes defined by slope gradient and length are indicated by coloured area symbols. The genetic terrain forms, and the past and present processes which cause them are listed in the legend where they can be determined on the basis of a hachuring system. In the Belgian system, morphometric information forms the basis for mapping, and slope steepness classes are indicated by coloured area symbols. One may, with justification, say that too much emphasis is thus laid on an aspect that certainly is not the essence of geomorphology. As a consequence insufficient means of cartographic expression are left for genetic and chronologic data. Morphological mapping in the United Kingdom is based on precise mapping of slopes with special attention to slope discontinuities.

Applied geomorphological maps have always been emphasized at the International Institute for Aerial Survey and Earth Sciences (ITC), the Netherlands. The ITC System of Geomorphological Survey and Mapping (Verstappen 1970, Verstappen & Van Zuidam 1975) was devised as an analytical geomorphological survey method, and thus morphometric, morphographic, morphogenetic and morphochronologic aspects of

Systems / Elements	Belgium: Univ. of Liège	Belgium: Univ. of Leuven	The Netherlands	Federal Republic of Germany	France	Unified key	ITC
Morphometry[1]							
classes of slopes	shades of colour	black stripes perpendicular to the contour lines	white line screens[2]	grey line screens		shades of colour	
break of slopes	black symbols		black and brown symbols	black symbols	coloured symbols[3]	symbols in full colour	black symbols
curvatures	black symbols			grey lines			only water sheds in black line symbols
Morphography	black symbols	black and red symbols	[2]	black symbols	coloured symbols[3]	symbols in full colour	
Morphogenesis	coloured surfaces; minor forms in black karst in purple symbols		coloured surfaces + cross-reference in legend	coloured surfaces	colour	coloured surfaces[4]	coloured surfaces
Morphochronology	black characters	the shades of the coloured surfaces include the age	cross-reference in legend	black symbols for inactive, orange-red signatures for recent processes; shades in coloured surfaces	shades of colour[3]	black characters	black characters
Hydrography	blue from base map		shade and thickness of blue lines indicate position below or above the surrounding area and width of water course	blue symbols	blue from base map	blue from base map	blue from base map
Geological structure			some symbols in red can be found in: "other symbols used"	locally strike and dip symbols in black + separate coloured map in scales up to 1:500000	colour	orange-red colour	black symbols
Lithology	black symbols + coloured map 1:160.000		[2]	brown-red symbols	coloured symbols[3]	symbols in full colour	grey symbols

[1] All maps should be printed on base maps with contour lines in grey or brown.
[2] All units in the Dutch system are mainly determined by differences in local relief and morphographic form groups, lithology and stratigraphy may also be incorporated within the units.
[3] Colours are determined by the morphogenetic process or geological structure.
[4] As one of the genetic types of landform the endogenous landforms are considered.

Table 1 Comparison of the contents of maps in several systems (from Salome & Dorsser 1985).

landforms are shown. Also both forms and processes, as well as geological structure and lithology, are included. In other words, an analytical geomorphological map must give information about the appearance (morphology), the dimensions and slope values (morphometry), the material (geology), the origin (morphogenesis) and the age (morphochronology) of each landform.

From this it may be clear that a general system cannot provide so much detailed information (Table 4.1). If a system is flexible, more possibilities exist to adapt to the special local differences in landforms. Therefore, the approach developed by the ITC System served as a guide line for comparable types of studies. This approach is adhered to because it is strongly felt that the genetic and chronologic characteristics of landforms are of great scientific and practical value in an investigation of the influence of neotectonism on geomorphic development within the study area. In this way map-making is not hampered by a limited choice of colours and map legibility is improved.

The basic problem with geomorphological maps is the great variety of information that could be included. Restraint in this respect is an absolute necessity; otherwise the maps produced will be complex, costly in printing and difficult for the user to read. Of the various legends proposed for detailed geomorphological mapping, very few meet all the desirable requirements (Table 4.2). The use of many symbols, such as coloured area symbols, line symbols (including hachures and screens of all types), numbers, letters or indices will have to be used to full advantage in order to produce a satisfactory map that is not overcrowded. Simplicity, by emphasizing the most essential information and generalizing or omitting less important

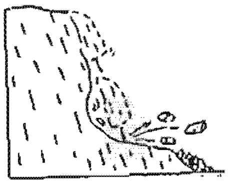
Systems / Colours and symbols	Belgium Univ. of Liège	Univ. of Leuven	The Netherlands	Federal Republic of Germany	France	Unified key	ITC
Colours	genesis		genesis	genesis	genesis and geological structure	genesis (endogeneous landforms included)	landform unit (genetic or structural)
Coloured surfaces	genesis		genesis	genesis		genesis	landform units
Colour shades	classes of slopes	genesis/ chronology	relief	usually: chronology[a]	chronology	classes of slopes	
Coloured symbols	purple for karst features	red for morphography	brown symbols incorporated in some units, red and brown for some 'other symbols'	orange-red for recent processes, brown-red for lithology	only symbols are used	usually symbols are printed in full colour	
Black symbols	morphometry morphography lithology minor genetic forms		incorporated in some units and in 'other symbols'	morphography, minor landforms, inactive processes, supplementary information			morphometry
Black characters	chronology		code of unit	supplementary information		chronology	chronology
Grey symbols[b]	influence of geological structure on landforms[c]			morphometry, some symbols with supplementary information			lithology
Striped colours			in some cases: landforms in built-up areas	polygenesis			
Mixed coloured symbols					polygenesis	polygenesis	
Striped coloured Symbols				alternation of sub-surface material			

[a] According to the legend colour shades may be used for:
- differences in accumulation and denudation
- differences in denudation
- differences in genesis
- differences in geological structure
- differences in chronology

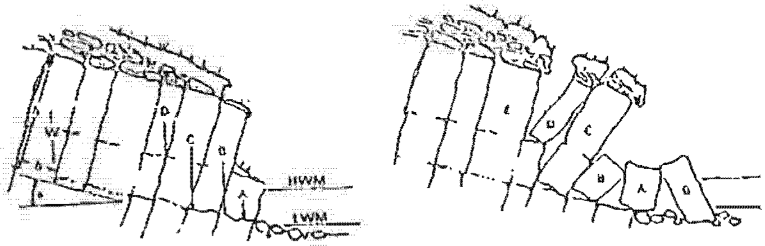
[b] base maps are usually printed in grey

[c] mentioned in legend, not used in the published sheets.

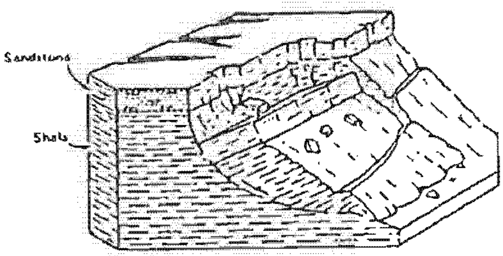
Table 2 Comparison of the use of colours and symbols in several systems (from Salome & Dorsser 1985).



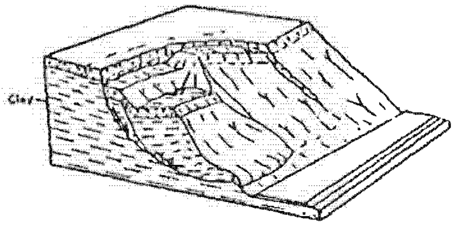
(a) Rock fall



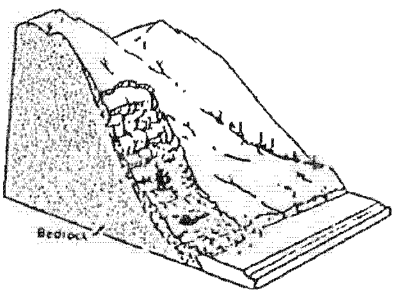
(b) Toppling failure



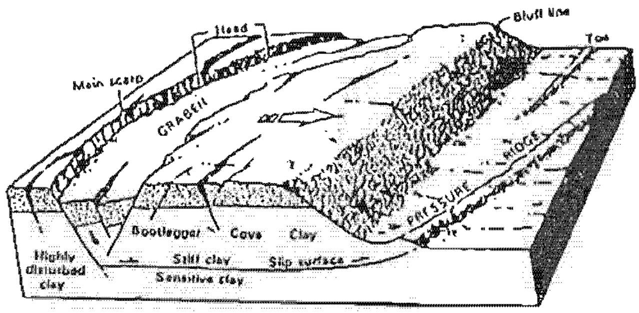
(c) Rotational block slide



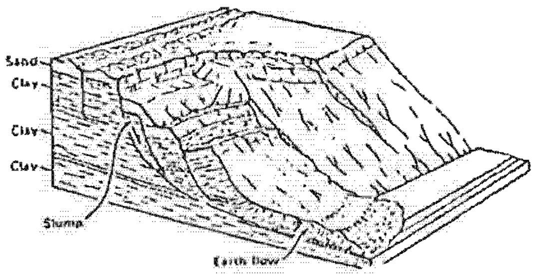
(d) Rotational earth slide



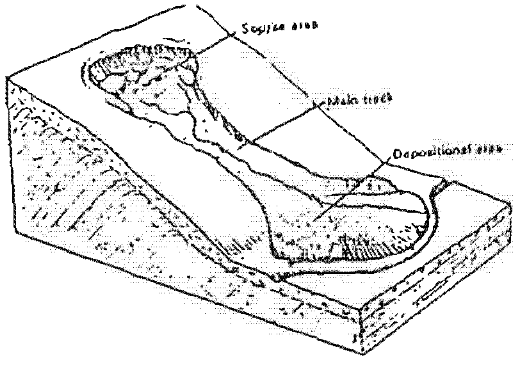
(e) Translational debris slide



(f) Translational block slide



(g) Slump-earth flow



(h) Debris flow

Figure 4.1
Principal slope failure mechanisms and modes as illustrated by Varnes (1978, fig. 2.1).

information, should be a guiding principle. If all information is considered indispensable, the data may be divided in an appropriate way over two or more maps covering the same area.

Mass movement classification is an attempt to categorise material that has moved downslope, usually classed according to similar morphology or genesis. Mass movements exhibit great variety, being affected by geology, climate and topography, and their rigorous classification is hardly possible (Fig. 4.1). Here the various types are subdivided mainly by mechanism, insofar as this is known, and morphology. Consideration is also given to the rate of the movement and the nature of the material involved. The Varnes (1978) terminology for slope stability problems has been adopted for this study due to its emphasis on description rather than categorization of a slope failure. Principal criteria used in classification are:

1. Mechanism and mode of movement
2. Type of material involved

Movement between bodies is relative and therefore refers only to the materials in relative motion, i.e. a slide refers to motion between stable and moving ground, however the term flow refers to the distribution and relative movement of particles within a moving mass. Falls comprise the more or less free descent of masses of soil or rock of any size from steep slopes or cliffs.

The adopted graphical presentation of the various forms and processes are the same ones as those in the ITC system (Verstappen & Van Zuidam 1968) with some modification to suit the needs of this study.

4.1.4 INTRODUCTION TO THE RANGE OF LANDFORMS IN THE STUDY AREA

Geomorphic studies provide a picture of the actual landscape of the Waipara region. They can suggest geomorphic mechanisms that led to the present situation and even provide a history of the course of landscape evolution. The gross morphotectonic situation has been summarised (see Chapter Three) and indicates that most of the folds and faults are arranged in a southwest trending dextral en echelon system. Nevertheless, a morphotectonic study is not complete without some speculation on fundamental causes of landscape evolution and the geomorphic evidence can be used to constrain and evaluate competing hypotheses.

Differential erosion of folded, interbedded resistant conglomerates, sandstones and limestones and less resistant mudstones and siltstones has produced a series of alternating homoclinal ridges and valleys. The en echelon folds of the coastal range illustrate well the variations in topography produced by folded rocks. For example, the rocks of Kate Anticline on the edge of the shear zone were folded in the late Pleistocene time and erosion has not yet begun to breach the crest of the fold. By contrast, the more central Cass, Montserrat and Black Anticlines developed in late Pliocene, and show steep or inverted northwest limbs, locally cut by high-angle reverse faults along which older basement rocks of the core to abut middle and late Tertiary sedimentary rocks. With these structures topography, while strongly influenced by bedrock control, is not a simple reflection of the structural relief. The anticlines have undergone deep incision, and now resistant units form homoclinal ridges corresponding to the strike of the limbs of the folds (Stereomodel 8&9). Homoclinal ridges converge in the direction of plunge of anticlines, but diverge in the direction of plunge of synclines.

The relationships suggest that continued subaerial weathering and erosion aided by concurrent tectonic uplift of the land combined with eustatic sea level change has produced a structurally controlled landscape. Such processes may have gone on since the Early Pleistocene. During this span of perhaps two million years or more, the drainage has been continually adjusting to changes in the lithologic and structural character of the bedrock as the ground surface has been lowered.

Throughout the Quaternary, sea level fluctuations of up to 150 m have occurred in response to the build up of continental ice sheets and their consequent melting (Shackleton and Opdyke 1973). An extensive flight of uplifted Pleistocene marine terraces occurs at several localities on the eastern slopes of the coastal range of the Waipara region, and have been mapped in detail by Carr (1970). The difference in height between terraces and the present shoreline represents their net uplift. Determination of uplift for marine terraces permits the description of tectonic uplift patterns through much of the late Pleistocene, which are discussed in the next chapter.

Most rivers and creeks whose courses are, or have been graded enough to allow for periodic floodplain development, display suites of terraces ranging in size from barely discernable flat patches to extensive plains many kilometres long and about a kilmetre wide. Research has been carried out in the Omihi (Harris 1982), Waipara (Wilson 1963), lower Waipara (Campbell & Yousif 1985), and Rakaia (Soons 1963) River valleys, and the general conclusion is that fluctuations in climate during the later Pleistocene were of sufficient magnitude to induce changes in the regime of the rivers that would allow for periodic aggradation and degradation, on a scale large enough to produce terracing.

During the Otira Glaciation, the climate was sufficiently cold to produce large volumes of debris in the Southern Alps by frost-shattering. Not only did this debris mantle the slopes and smother the valley heads but it also choked the river valleys, mainly as a result of a declining load-discharge ratio brought about by the less humid conditions. Once the climate ameliorated enough to produce greater precipitation and resultant channel flow, the alluvium in the channel was removed downstream, and subsequently rivers entrenched their own deposits and even the underlying bedrock. These processes, of periodic aggradation and degradation in the upper and lower valleys, occurred several times after the formation of the Teviotdale Surface (Plate 2), with the result that most valleys exhibit suites of terraces reflecting their degradation origin (Campbell & Yousif 1985). Wilson (1963) correlated the Teviotdale Surface elsewhere with the depositional surface that was formed during the penultimate advance of the Otira Glaciation. During the post-glacial period the river have incised within the earlier deposits, formed a Holocene river terrace .

A common feature of river valleys in tectonic areas is an increase in river terrace gradient with increasing age Yousif (1986). The age and height distribution of fluvial surfaces crossing active folds and faults throughout the study area is investigated as part of this study to provide data on continuing activity during the late Quaternary (Holocene) and this is discussed in Chapter Six. The fluvial deposits of the Waipara area are all topographicaly lower than the highest marine terrace of the last interglacial sea level maximum, based on field observations, this is discussed in detail in Chapter Five.

The present drainage system was initiated as consequent upon the original surfaces subsequently affected by the main movements of the Kaikoura Orogeny. On the other hand, this drainage pattern once established is not easily disrupted by tectonic movements because the rivers are entrenched in deep gorges and they tend to maintain their courses throughout the tectonic evolution (antecedent). Consequently, the drainage pattern may be used to trace this evolution from the early stages. Several pieces of evidence suggest that several rivers in the region are antecedent. The Waipara and Teviotdale Rivers together with Kate, Carrington and Yellow Rose Creeks provide good examples of an antecedent origin which are discussed in Chapter Six. Antecedent rivers and their valleys are common only in actively orogenic regions. During deformation antecedent streams may maintain their valleys across growing folds and drainage networks are likely to develop with their long axes roughly normal to the coast, or transverse across orogenic belts. Later, as subsequent valleys deepen and lengthen, regional drainage networks are likely to develop trellis patterns, parallel to the structural trend (in this case parallel to the coast) which are discussed in Chapter Seven.

4.2 PRINCIPLES OF GEOMORPHIC CARTOGRAPHIC DESIGN ADOPTED FOR THIS STUDY

4.2.1 BASIC PRINCIPLES

The mapping system adopted in this study provides a means of establishing straightforward relationships and highlights anomalies which may throw light on interesting and even important aspects of geomorphic evolution.

Regionalization of terrain has long formed an important component of terrain analysis and continues to be a powerful tool in several distinctive ways. Such regionalization can use data from topographic and thematic maps and ground survey, but historically has relied heavily on conventional black and white air photographs and nowadays increasingly on the newer forms of remote sensing imagery (Peplies & Keuper 1975).

Aerial photographic terrain analysis is the systematic study of visual elements relating to the origin, geomorphic history, and composition of distinct landscape units that appear in aerial photographs. Through the analysis of pattern elements visually apparent on an aerial photograph, the geomorphic composition or parent material of a site is interpreted or inferred (Way 1973). These patterns are composed of several major elements that are evaluated by interpretation, for example: topography, drainage pattern, gully characteristics, erosional features, landform boundaries, tone, texture, vegetation and any other special features that may be present.

In this study a methodology which is essentially physiographic has been used. Such an approach inherently implies that the land surface can be readily divided into a finite number of units having natural boundaries. Furthermore, the assumption is made that analogous units may be recognized, which enables prediction between these units. The methodology is similar to that defined by Zuidam & Zuidam (1979). Their approach to the study of the various terrain forms and units is through the construction of cross-sections or profiles of the terrain showing topographic, geologic, geomorphologic and pedologic characteristics (Zuidam & Zuidam 1979, p.11). In this study the evolution of the major landform assemblages will be described with reference to (Fig. 4.2). Three landform assemblages are recognized in the basic model, each consisting of convex-concave or convex-rectilinear elements. Each landform assemblage reflects a different balance between the various processes of landscape denudation, structure and accumulation, therefore the model may be of benefit in the direct study of these processes as it highlights the relative degree of geomorphological activity of the landscape components.

The limits of the mapping evaluation depend on predominant characteristics present on the ground and visible on air photographs, which can develop only when sufficient outcrop and the rock has some uniformity in dip. Only under such conditions will a typical drainage pattern develop and furnish clues for identification. The basic points of the mapping are as follows:

1. Morphographic and morphometric aspects are represented by basic relief lines (break of slope) which divide the relief into unit slopes. This approach is based on the assumption that the land surface is composed of units which are either convex-concave or convex-rectilinear in profile with slope discontinuities bounding them. This is based on the concept that most of the land surface in the hilly and mountainous land is composed of individual slopes. A break of slope occurs where there is a sharp junction line between two slopes, or morphological units of differing steepness. These forms can be identified and delineated from the aerial photographs of 1:25,000 scale during the preliminary analysis. Field checks can be used to determine the accuracy of the interpretation and to gather more field data for geomorphological evaluation during the operational stage. These two stages are discussed in Chapter One.

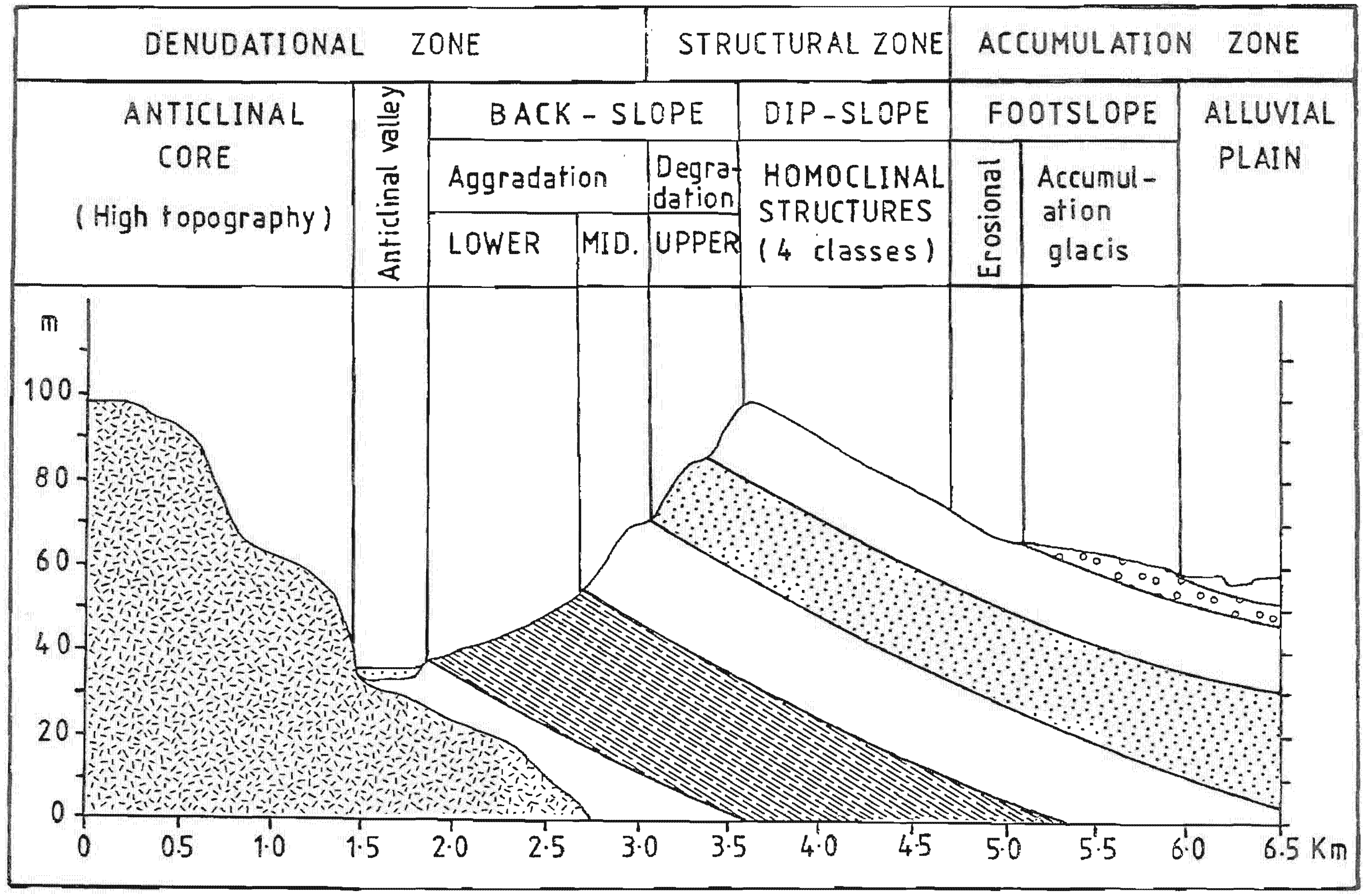


Figure 4.2 Example of a cross-section which cons titude the framework for terrain analysis and classification showing topography, geology, and geomorphic characteristic.

A topographic (contour) map gives much essential information in this respect and thus is printed on a separate overlay map.

2. Materials which constitute the surface terrain are divided into 4 classes. The types and features of the sediments may be interpreted from the genetic classification of a landform and/ or distribution of geomorphological processes. The rock units are grouped according to broad lithologic types having different resistance to erosion. For example:

a. units containing extensive outcrop areas of resistant rock that tend to form highlands or high relief; e.g. homoclinal structures on Amuri Limestone, Weka Pass Stone etc and Torlesse Supergroup terrain.

b. units containing alternating sequences of resistant and non-resistant lithologies, which, where tilted or folded, produce elongate ridges and valleys or a distinctive dip and scarp topography; e.g. homoclinal structures on Greenwood Fomration, Waikari Formation, Mt Brown Formtion etc.

c. units containing mostly non-resistant lithologies, and hence are eroded to form topographic depressions; e.g. anticlinal valleys on Conway Formation.

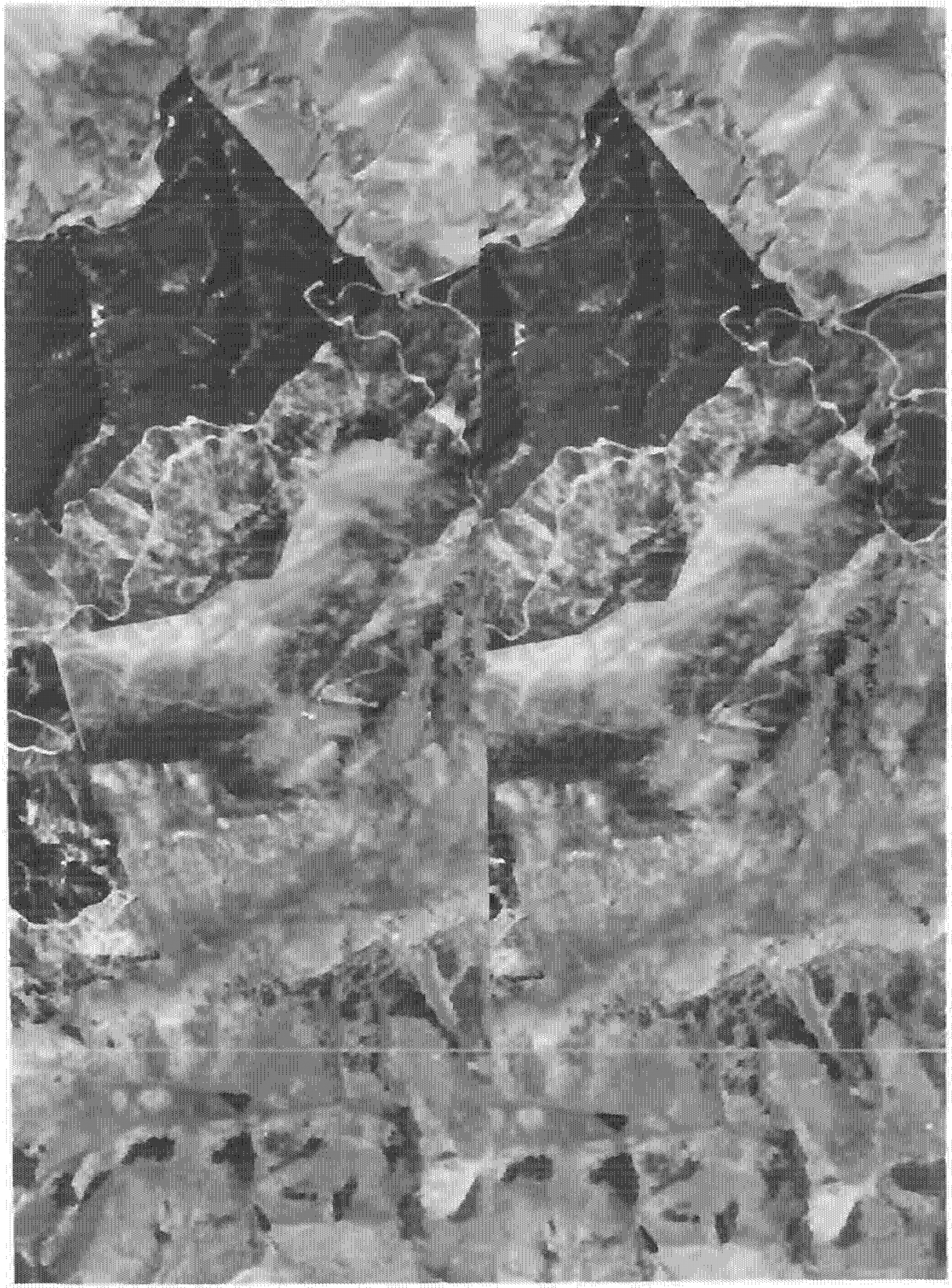
d. units occupying gently sloping or flat plains; e.g. alluvial plains (Canterbury and Teviotdale Surfaces) and marine surfaces.

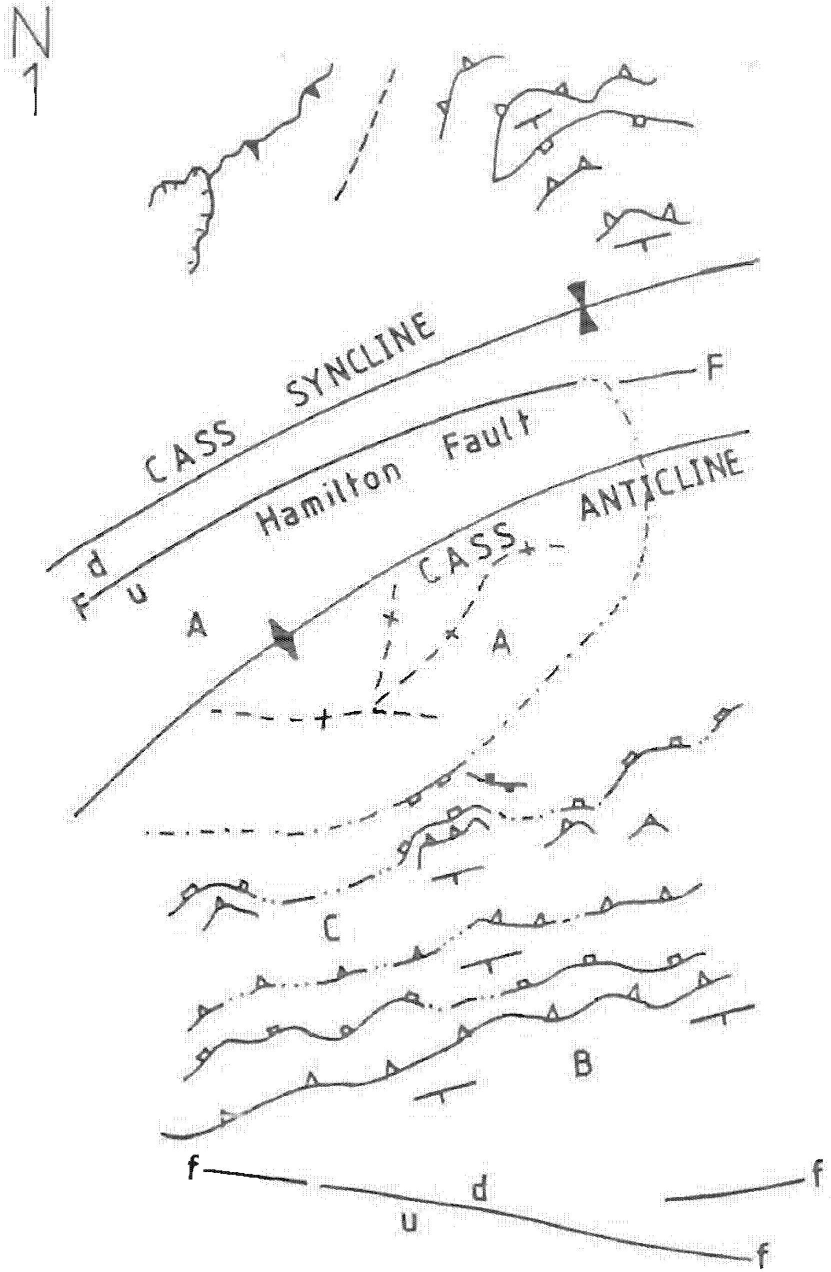
e. units occupied by the river bed and surrounding floodplain and terraces.

Interpretation of aerial photographs for terrain classification relies upon drainage pattern/texture identification and analysis. These are used as general indicators of rock properties. Because drainage is particularly sensitive to structural trends, a detailed

Stereomodel 8

Stereomodel of a maturely dissected, folded and faulted landscape. The Cass Anticline is an unroofed fold exposing both the basement (area A) and the covering strata. Only the southern limb is found in this area as the northern limb is completely faulted out by Hamilton Fault. Note the major and minor fault traces.




N
CASS SYNCLINE
Hamilton Fault
CASS ANTICLINE
F
F
d
u
A
A
C
B
f
f
f
d
u

drainage map of the study area was prepared and used as a base for plotting structural, geomorphological and lithological data.

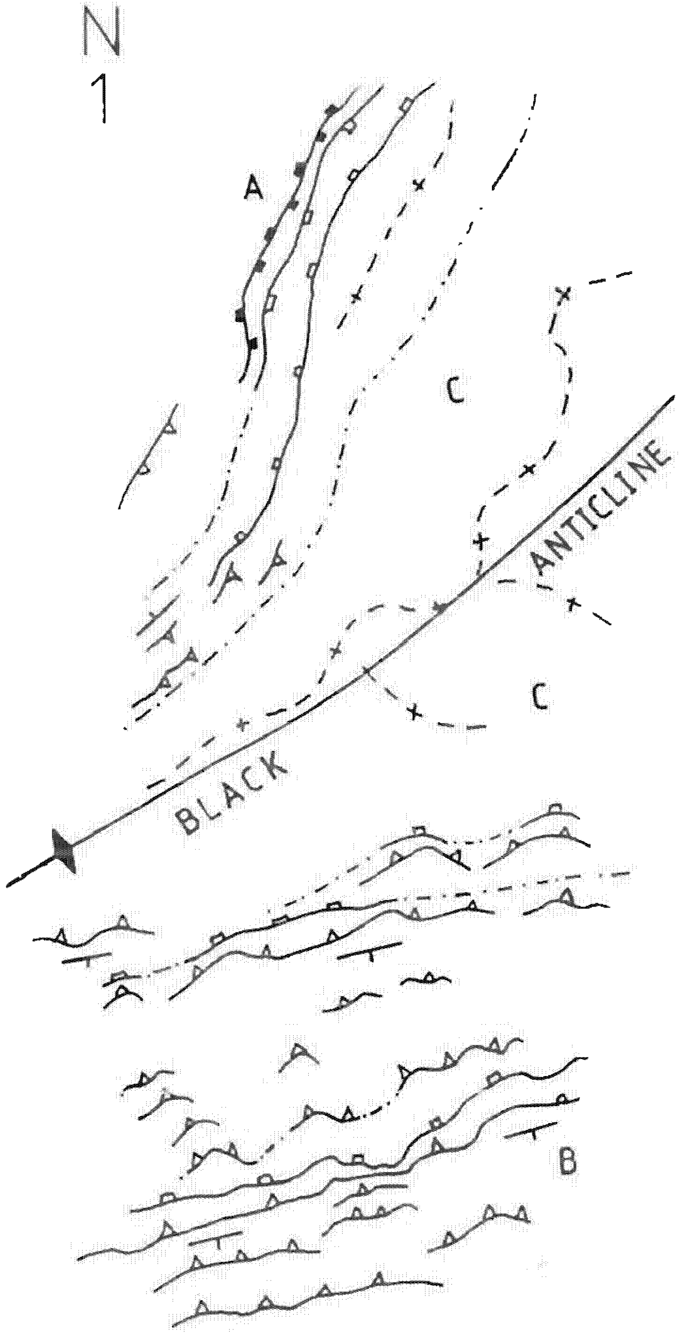
3. Genetic classifications of structural, denudational, fluvial, marine, etc. landform units are mapped on the main map, and areal distribution of geomorphic processes is shown also in the same map. Delination of several erosional processes and forms helped to a great extent in such classification. The best means of cartographic emphasis, coloured area symbols, are used for major genetic landform types, but not for lithology and chronology as has been used by other systems discussed before.

The distinction between areas of degradation and areas of aggradation has been used as a major distinction along the backslope of homoclinal structures, indicated by differences in coloured area symbols. Although the above distinction is important, the cartographic presentation has the disadvantage that morphogenetically very different low or high land areas are put in the same category. It may also be argued that rock type, although geomorphically important, is overemphasized in this way. To fill out this scheme a further subdivision is made on the basis of lithology, subaerial processes, and relief amplitude.

4. Actual processes and their geomorphic expression, e.g. patterned ground, are mapped on the basis of aerial photograph interpretation and verified by a detailed field check. These are indicated by line symbols for which only the black colour is used so as to simplify the mapping system. The cartographic elaboration of the map should allow for the identification and exact description of the landforms and landform complexes and should indicate their position and arrangement as well as their genesis.

Stereomodel 9

The stereomodel reveals a structurally controlled landscape. The Black Anticline is an unroofed fold exposing both the basement and the covering strata. The dike like ridges (area A) and hogbacks (area B) of the outer flank of this anticline are composed of Amuri Limestone and Waikari Formation respectively. The iner dome core (area C) is composed in Torlesse Supergroup. The subsequent valley is cut along a soft bed of Conway Formation and Loburn Mudstone.


N
A
C
C
BLACK ANTICLINE
B

5. In this study chronological information is the core of the mapping, but it was considered of secondary importance, and priority was given to the morphogenetic landform (origin) units for which purpose the colours were used. Therefore ciphers are used to indicate chronology. The advantage of this is that if the age of certain landforms is unknown, the cipher is simply omitted. It seems advisable to use this method, since in most areas chronology is the most difficult and subjective information to determine, and also susceptible to modification by the introduction of new data.

4.2.2 CLASSIFICATION

A. Units of Structural Origin

The present distribution of homoclinal structures in the landscape reflects the role of structure and lithology in their morphological pattern which can be divided into four classes on the basis of angle of dips and their photo patterns. Well known "homoclinal" structures include cuestas, hogbacks and dike-like ridges (Zuidam & Zuidam 1979). The landscape thus formed are classified as dike-like ridges, hogbacks, cuestas and the dissected form of the homoclinal structures "flatirons" topography (Fig. 4.3).

i. dike-like ridges

This unit occupies a number of sinuous ridges on the north-western limb of the Cass, Black and Montserrat Anticlines where they are associated with the steep limbs (Stereomodel 8&9). The dike like ridges may be formed by near vertical dipping rock. They have narrow, steep slopes, and are basically symmetrical with very steep concave slopes which may be either smooth or irregular. This unit has

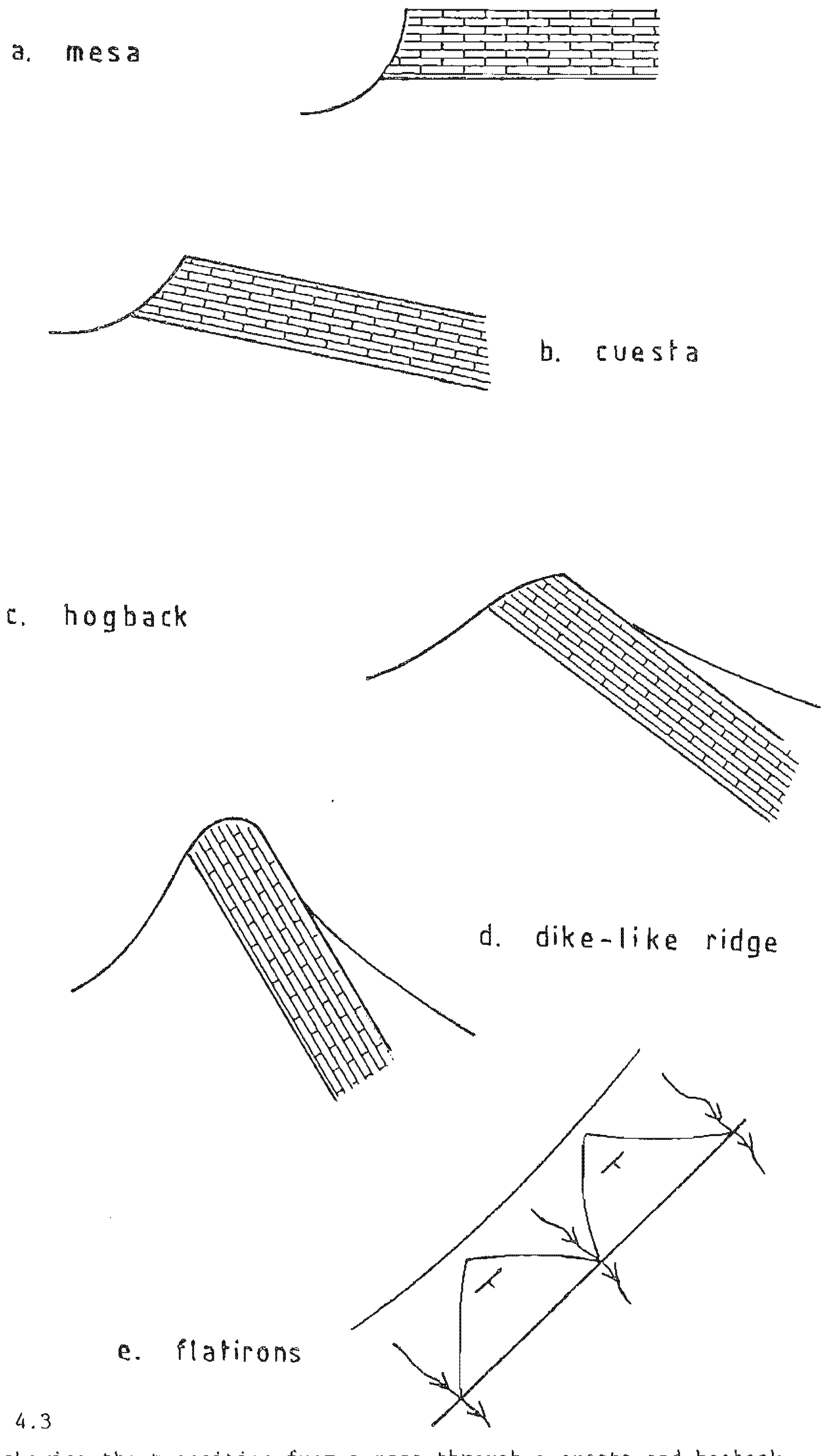


Figure 4.3
Diagram showing the transition from a mesa through a cuesta and hogback to a dike-like ridge as controlled by varying rock dip (modified after Davis 1902).

a parallel drainage pattern.

ii. hogback ridges

This unit consists of elongated parallel ridges and steep slopes that merge into footslopes which may be either convex or concave and smooth or irregular in form. Hogback ridges form the dominant structural units in the area, where the outcrop of resistant ridge-making sandstones, limestones and conglomerates is controlled by folding and faulting (Stereomodel 8).

Mass movement closely associated with these steep scarps and dipslopes include rockfalls, slumps, translational slides, earth flows, and rill and gully erosion.

iii. cuestas

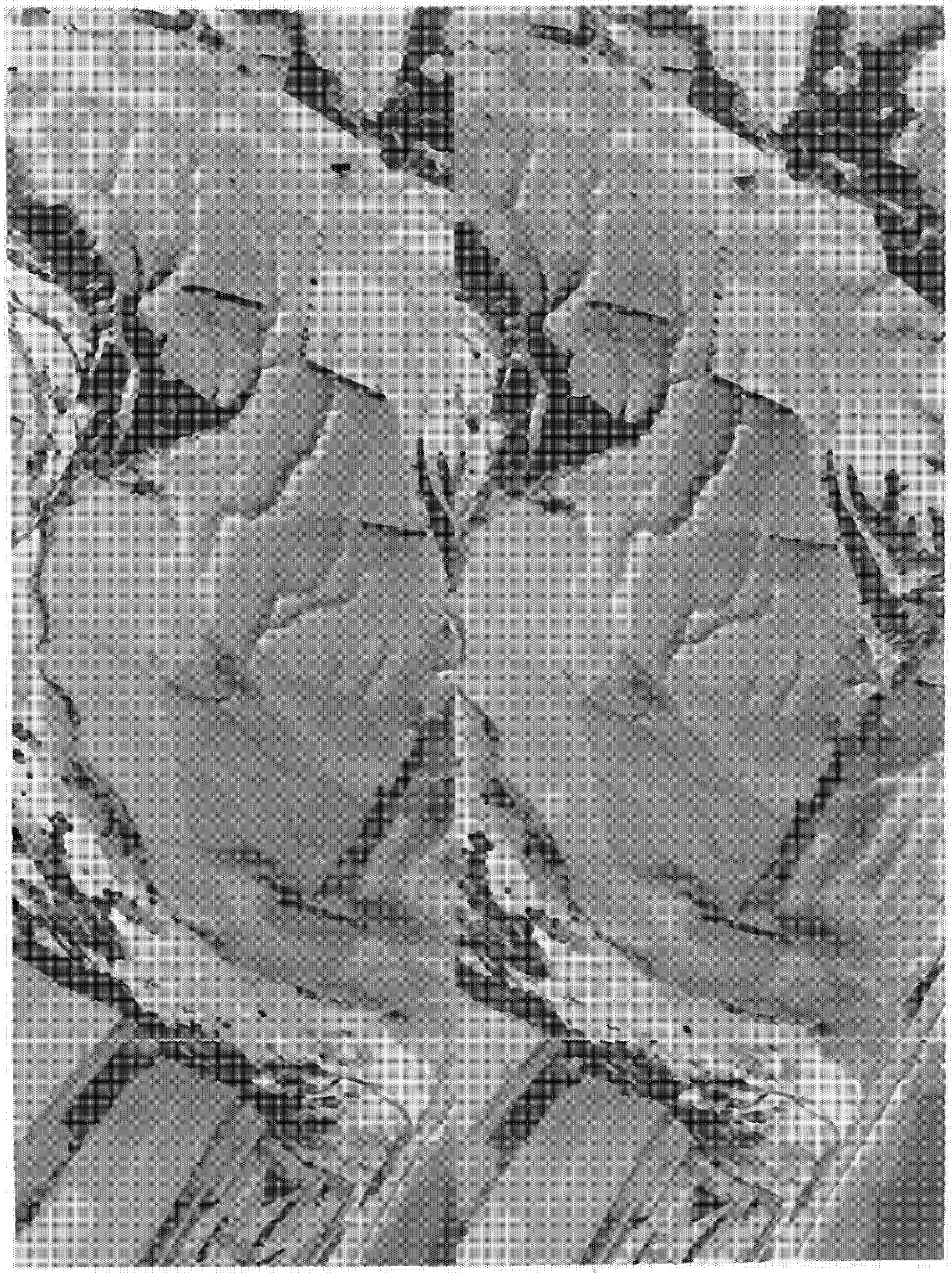
These form on the gently to moderately dipping slopes of the folded landscape. This unit is typically composed of an asymmetrical ridge with one long and gentle slope generally coinciding with the dip of the resistant bed or beds and one short, steep slope or scarp. The former is smooth and may have a dendritic or parallel drainage pattern, while the steep slope is typically concave and irregular in form (Stereomodel 10).

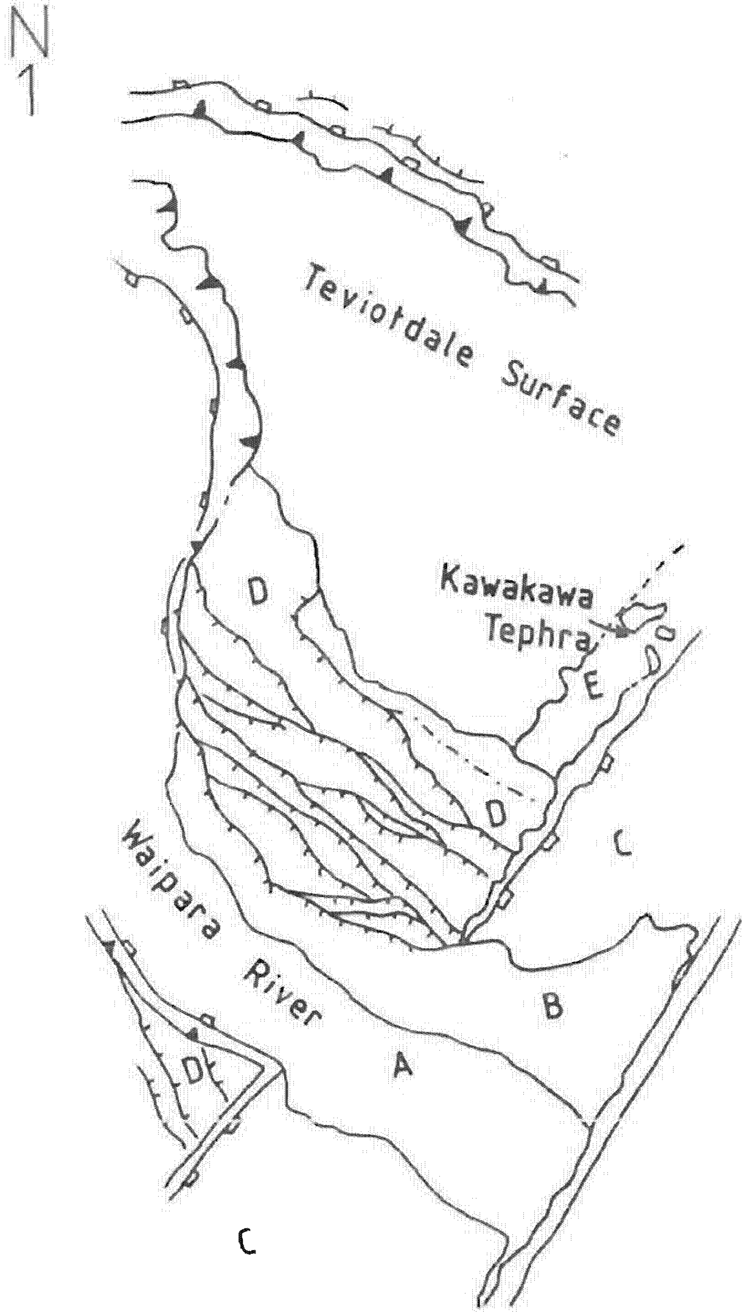
iv. flatirons

Triangular shaped homoclinal structures are called flatirons because they resemble the old-fashioned iron resting on its base (Scott et al. 1985). This unit type has a triangular shaped, steep dip slope which is fringed on its side by a narrow edge or ridge crest. Below the ridge crest are short, very steep scarps. In

Stereomodel 10

Distinctive landform of fluvial and marine terraces near Waipara River mouth. The step like landform of fluvial terraces is conspicuous (area D). Note active and inactive flood plain (area A and B respectively). Note also the Holocene coastal plain (area C), the uplifted marine terrace (area E), and the coastal cliff.





N
Teviotdale Surface
Kawakawa
Tephra
D
E
D
C
Waipara River
B
A
D
C

practice, this unit is confined to the south-eastern limb of Montserrat Anticline and is thus of limited application to the area. Flatirons may occur singly, or as a series stacked one upon another (Stereomodel 11).

Table 4.3: Lithologic subunits of structural landforms

a. Homoclinal structures on:

1. Greenwood Formation (SGF)
2. Mt Brown Formation (SMF)
3. Greta Formation (SGF)
4. Waikari Formation (SWF)
5. Omihi Formation (SOF)
6. Amuri Limestone (SAL)
7. Broken River Formation (SBF)
8. Kowai Gravels (SKG)

b. Anticlinal cores, high rugged topography on Torlesse Supergroup (ST)

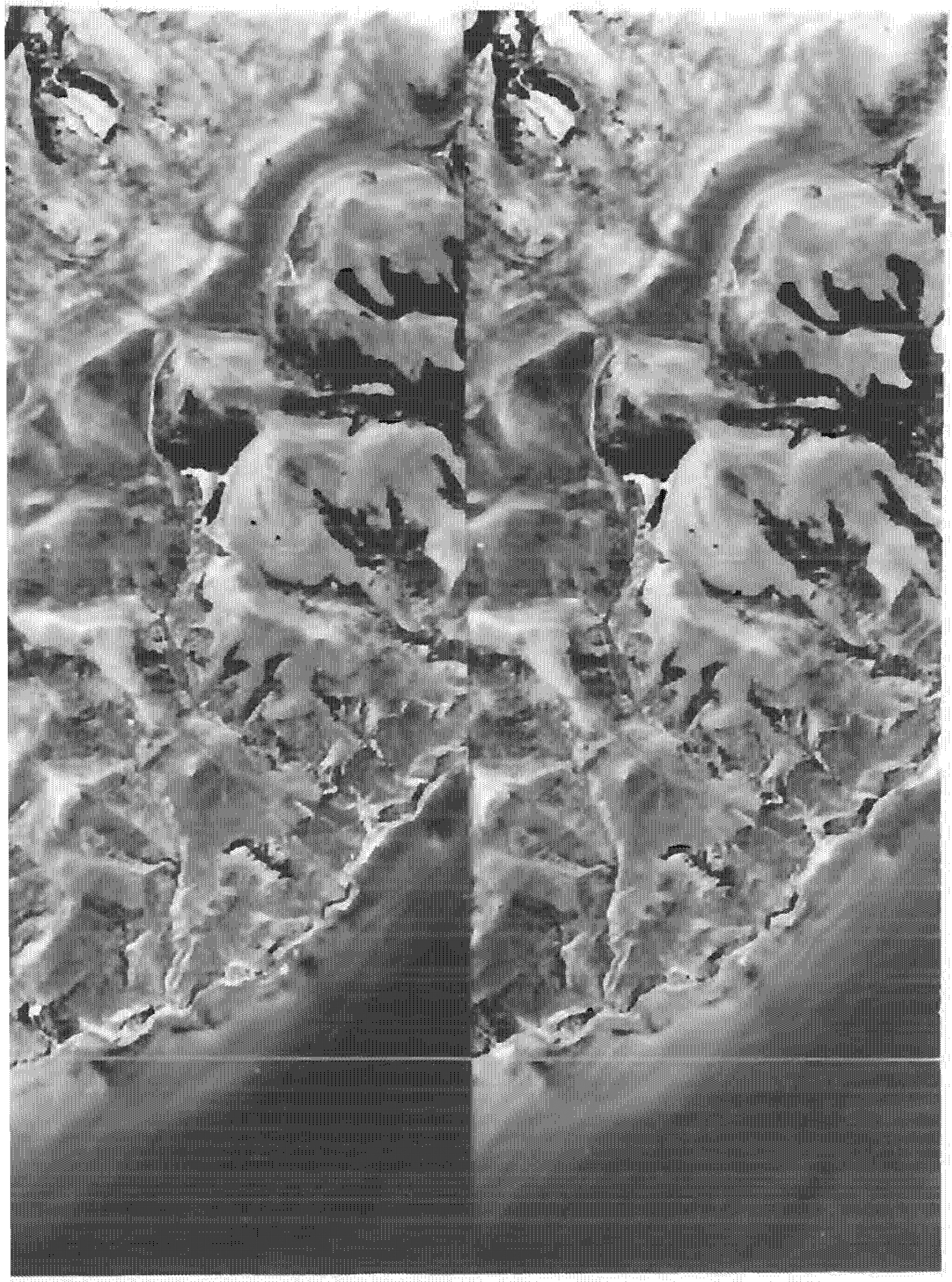
B. Units of Denudational Origin

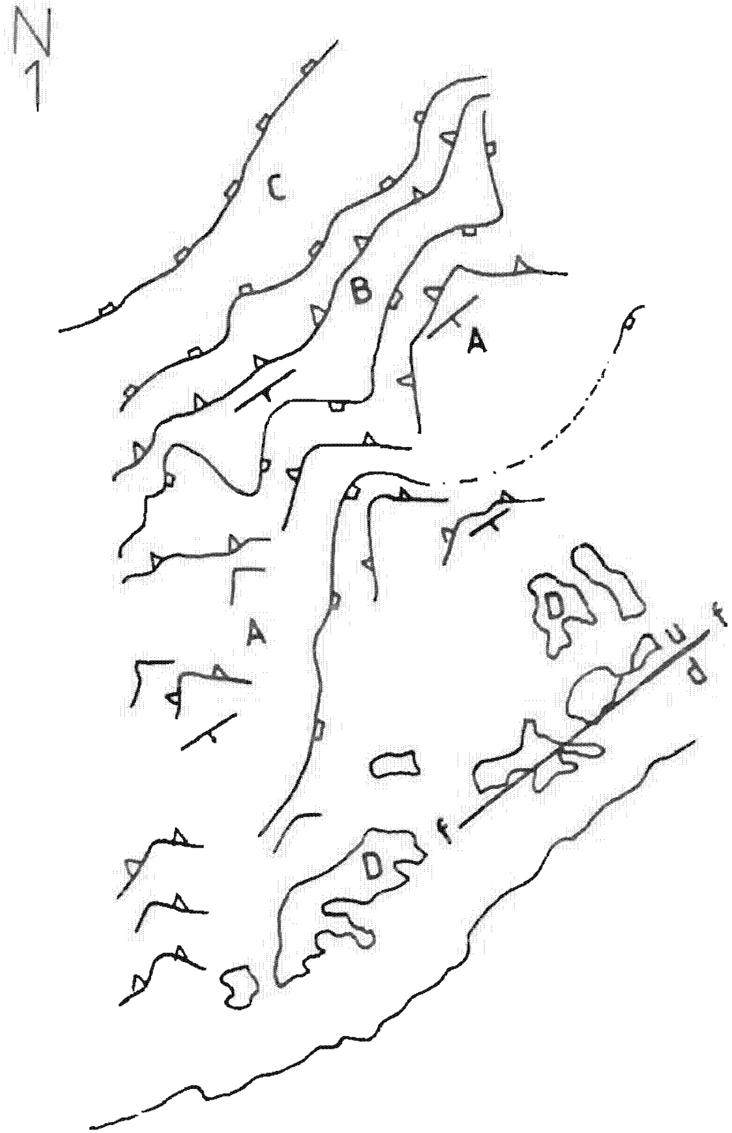
i. erosion glacis (footslope)

Erosional footslopes are landforms which are easily detectable on aerial photographs because of their gently inclined erosion surfaces. They are carved in bedrock by weathering and erosional processes and are generally veneered with debris. Erosion in the form of sheet erosion is predominant (Zuidam & Zuidam 1979). These surfaces slope gently away from the hills. Slope length varies from a few tens to thousands of metres. Gradients vary between localities, but all are characterised by ridges and furrows, as shown on the plate

Stereomodel 11

Stereomodel of a sequence of differentially eroded south east-dipping sedimentary strata and dissected, uplifted, marine terraces. The differential erosion of these units in zone A,B and C have produced numerous escarpments, hogback ridges, and flat iron topography. Note the uplifted marine terraces in the southern part of the stereoscopic area (area D). Note also the fault trace and the coastal landslides in this area.




N
C
B
A
A
D
D
u
d
f
f

2.

ii. accumulation glacis (footslope)

In the event that the amount of debris materials exceeds the amount of transported material, accumulation is predominant (Zuidam & Zuidam 1979). Usually it is believed that the climatic background for both erosion and accumulation glacis is the same and that an erosion glacis upslope serves as a source for the accumulation glacis, so that frequently one is found as a continuation of the other (Fig. 4.2). Accumulations, sometimes in the form of coalescing fans were recognized. Typically the surface is straight to concave, smoothly sloping and covered by unconsolidated Quaternary sediments. The author believes that sheet erosion and scarp retreat are the essential agent in forming the relief.

iii. backslope

This geomorphic unit occupies the backslope of the Cass, Black and Montserrat Anticlines, from the crest of the highest homoclinal structures downward to the anticlinal valleys (PLate 2).

The distinction between areas of degradation and areas of aggradation have been used as a main distinction along the backslope of homoclinal structures (Fig. 4.2). Three major units were recognised along the backslope.

iv. anticlinal valley

The eroded Cass, Black and Montserrat Anticlines are flanked by a well stratified resistant unit that forms distinctive dike-like and hogback facets. Long strike valleys have developed by headward erosion along the weak formations. This geomorphic unit, thus is developed parallel to the axis of the structure, bounded on both sides by homoclinal ridges (Stereomodel 8&9).

Table 4.4: Lithologic subunits of denudational origin

a. Undissected accumulation glacis (footslope) on Canterbury Gravels

b. Dissected accumulation glacis (footslope) on Greenwood Formation

c. Dissected erosional glacis (footslope) on:

1. Greenwood Formation (DGF)
2. Mt Brown Formation (DMF)
3. Waikari Formation (DWF)
4. Conway Formation (DCF)

d. Dissected upper backslope with structural control on:

1. Kowai Gravels (DKG)
2. Tokama Siltstone (DTS)
3. Amuri Limestone (DAL)
4. Ashley Mudstone (DAM)

e. Dissected middle backslope with structural control on:

1. Ashley Mudstone (DAM)
2. Waipara Greensand (DWG)
3. Greenwood Formation (DGF)

f. Dissected lower backslope with structural control on:

1. Loburn Mudstone (DLM)
2. Waikari Formation (DWF)

g. Anticlinal Valley

h. Erosional core of Black Anticline, low rugged topography on

Loburn Mudstone (DLM)

C. <u>Units of Fluvial Origin</u>

Extensive fluvial surfaces developed in response to the Late Quaternary glacial and interglacial periods. These are represented by multiple periods of aggradation, lateral channel migration, and the modern flood plain within large and small drainage basins.

i. alluvial plains

In the lower Waipara area between the foot hills of the Southern Alps and the sea there are two aggradation alluvial plains (Wilson 1963). The older, named the Teviotdale Surface, is now limited to residuals a few square kilometres in extent (Plate 2). Remnants of the upper surface up to 107 m above sea level and 50 m above the Waipara River, are best preserved on hills southeast of the Waipara river (Stereomodel 10). At lower altitudes than the Teviotdale Surface is an alluvial plain of better preserved and more continuous aggradation surfaces called the Canterbury Surface (Wilson 1963).

ii. floodplain and valley fill

Extensive floodplains are confined to the Waipara and Omihi valleys where they straddle the Waipara River and its tributaries across the Canterbury Surface (Stereomodel 4). Like most floodplains, they are characterised by low elevation (less than 1 m), it is composed of gravels, sands, and silts deposited during the present regime of the river, forming a variety of fluvial landforms.

iii. alluvial river terraces

Terraces are relatively flat, horizontal, or gently sloping surfaces, sometimes long and narrow, which are bounded by a steeper ascending slope on one side and by a steeper descending slope on the opposite side (Zuidam & Zuidam 1979). Where well-devoloped, they are typically step-like in character (Stereomodel 10). They are, in principle, old flood plains which are at the moment incised and are no longer flooded by the river. The older terraces are commonly dissected by gullies and valleys.

iv. gullies and rills

Gullies and rills are comparatively common on the dip slope and back slope of homoclinal structures. Shallow rills are developed in association with gullies. Good examples of these are to be found along Waikari Formation, Tokama Siltstone, Mt Brown Formation and marine terraces. Shallow rills are particularly well developed in Torlesse Supergroup terrain.

v. alluvial fans

Alluvial fans exist where a distinct boundary occurs between the coastal range and the surrounding plains. The fans vary considerably in morphology and extent, as a result of the characteristic of the catchment area and the local erosion base.

Table 4.5: Units of alluvial landforms adopted for this study

a. Upper aggradation surface (Teviotdale gravels) (FTG).

b. Lower aggradation surface (Canterbury gravels) (FCG).

c. Alluvial river terraces.

1. Modern river terraces (FMT).

2. Ancestral river terraces (FAT).

d. Flood plain and valley filling (FFP).

e. Major and minor gulley and rill and erosion.

f. Alluvial fans (FAF).

D. <u>Units of Marine Origin</u>

i. marine terraces

Almost flat surfaces gently sloping towards the sea. These terraces are progressively dissected by fluvial processes. The amount of fluvial dissection increases up the terrace flight, until the inner margins merge with the completely dissected interior region. Thus, these terraces can be classified into three types as shown on Plate 2.

ii. coastal cliff

A sea cliff is a high, steep face of rock formed by wave action. Such a cliff can be recognized on aerial photographs by its abrupt, steep and frequently linear appearance. Mass movement features closely associated with such steep cliffs include rockfalls, transtional slides and rill and gully erosion.

Table 4.6: <u>Units of landforms of marine origin adopted for this study</u>. Dating of the terraces is discussed in Chapter 5, section 5.7.

a. Isolated remnants of marine terraces

1. 125 ky marine terrace (M4)

2. 105 ky marine terrace (M3)

b. Dissected, continuous marine terraces

1. 80 ky marine terrace (M2)
2. 60 ky marine terrace (M1)

c. Young coastal plain with alluvial fans

1. Holocene marine terrace (M)

d. Steep, uplifted coastal cliff

e. Modern beach

E. Units originating from mass movement

i. Landslides

The term "landslides" has come to be a term covering all types of downslope movement and there are several confusing usages of the term in the literature. For example, Hutchinson (1968) defines landslides as "relatively rapid movements involving failure" which takes place on one or more discrete surfaces, but includes mud flows and falls under this heading. Varnes (1978) makes the suggestion that "landslides" be applied only to movements which involve a sliding mechanism and this is accepted by the author.

ii. rock falls

The term "fall" is applied by Varnes (1978) to the movement of the mass mostly through the air by free fall, leaping, bounding, or rolling from a steep slope or cliff, along a surface on which little or no shear displacement takes place.

4.3 REGIONAL DISTRIBUTION OF LANDFORM IN THE STUDY AREA

To describe the landscape of the study area and to attempt to explain its origin, it is convenient to divide the region into several physiographic units (Fig. 4.4). Although each landform terrain is topographically distinctive, these physiographic areas are distinguished by specific combinations of structural and lithological properties:

1. The eastern lowland includes the marine landforms.
2. The coastal range comprises highly deformed Torlesse basement and Cenozoic covering strata which, extends north-west from near Amberley to the northern margin of the area.
3. The central lowland is a fluvial aggradation plain forming the northern extension of the Canterbury Plains.
4. The northwestern highland includes Weka Pass moderately sloping terrain.

The coincidence of geomorphic expression and geology suggest a primary tectonic control on development of the pattern of landforms terrains. Present-day physiography is influenced by active fluvial and marine processes operating on the structural features produced by folding and faulting of the sedimentary cover by tectonic activity. Thus, morphology has been affected by geomorphological processes and by tectonic activity.

4.3.1 EASTERN LOWLAND

The coastal lowland comprises a number of surfaces which relate to episodes of Pleistocene planation and deposition and these are separated as formational units in the regional stratigraphy. They have already been discussed in detail in Chapter Two. Both marine and non-marine deposits were affected by subsequent geomorphic processes leading to separation into five classes as shown on Plate 2.

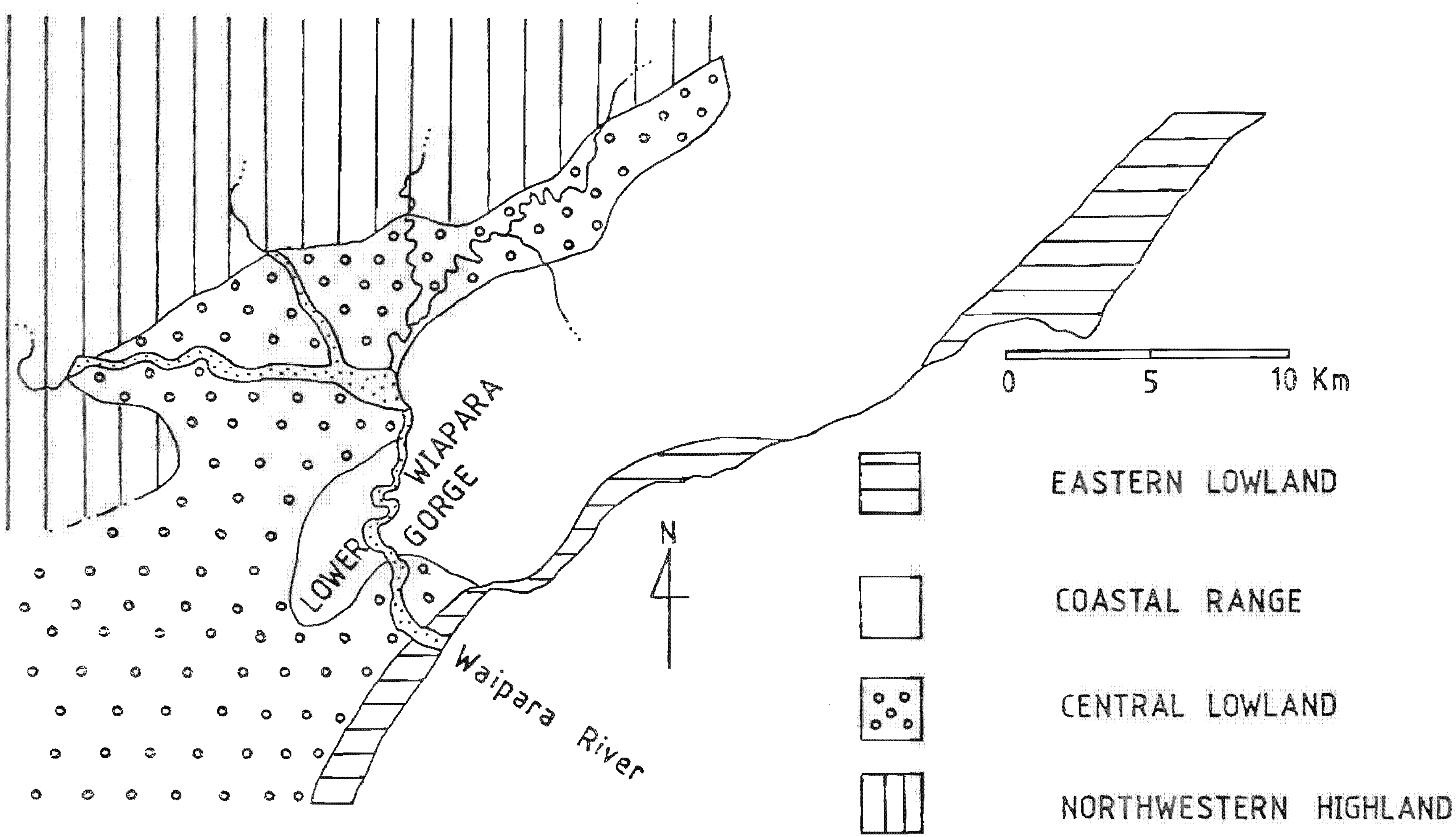


Figure 4.4

Distribution of the main physiographical units in the study area.

These marine surfaces can be traced south from Happy Valley to near the Waipara River with a reduction in elevation southward (Plate 2). Landforms on the upper surfaces are generally initiated before those of the lower ones and are more deeply dissected. The landforms range in width from about 0.5 to 3 km. Slopes of 5-15^{0} are characteristic of this terrain.

The present morphology of the coastal drainage is strongly affected by cyclic Late Quaternary eustatic sea level fluctuations overprinting differential uplift.

4.3.2 COASTAL RANGES

The coastal ranges extend north east from Amberley to the northern margin of the area, a distance of about 25 km (Fig. 4.4). The juxtaposition of resistant and weak formations has had profound effects on the geomorphology and erosional development of the ranges. The spectacular hogbacks and marked dip slopes caused by more resistant limestones, conglomerates and sandstones give interest and diversity to the landscape which is otherwise dominated by mudstone hills. The area includes the highest peaks as well as low hills, and some broad valleys (Plate 2). From an altitude of about 60 m at their southern end, the coastal ranges gradually rise north to the Cass and Black Anticlines, where many peaks reach altitudes of more than 520 m. The basement rocks of Upper Jurassic and Lower Cretaceous form the core of many anticlines. On the west side, the ranges gradually descend to the floor of the central lowland. In many places, no sharp boundary exists between the coastal ranges and the lowlands. Slope-wash deposits form the rounded junctions between the ranges and the lowlands. The present elevation is usually a product of Late

Tertiary and Quaternary uplift and deformation has continued to the present day. Further discussion is reserved for Chapter Six.

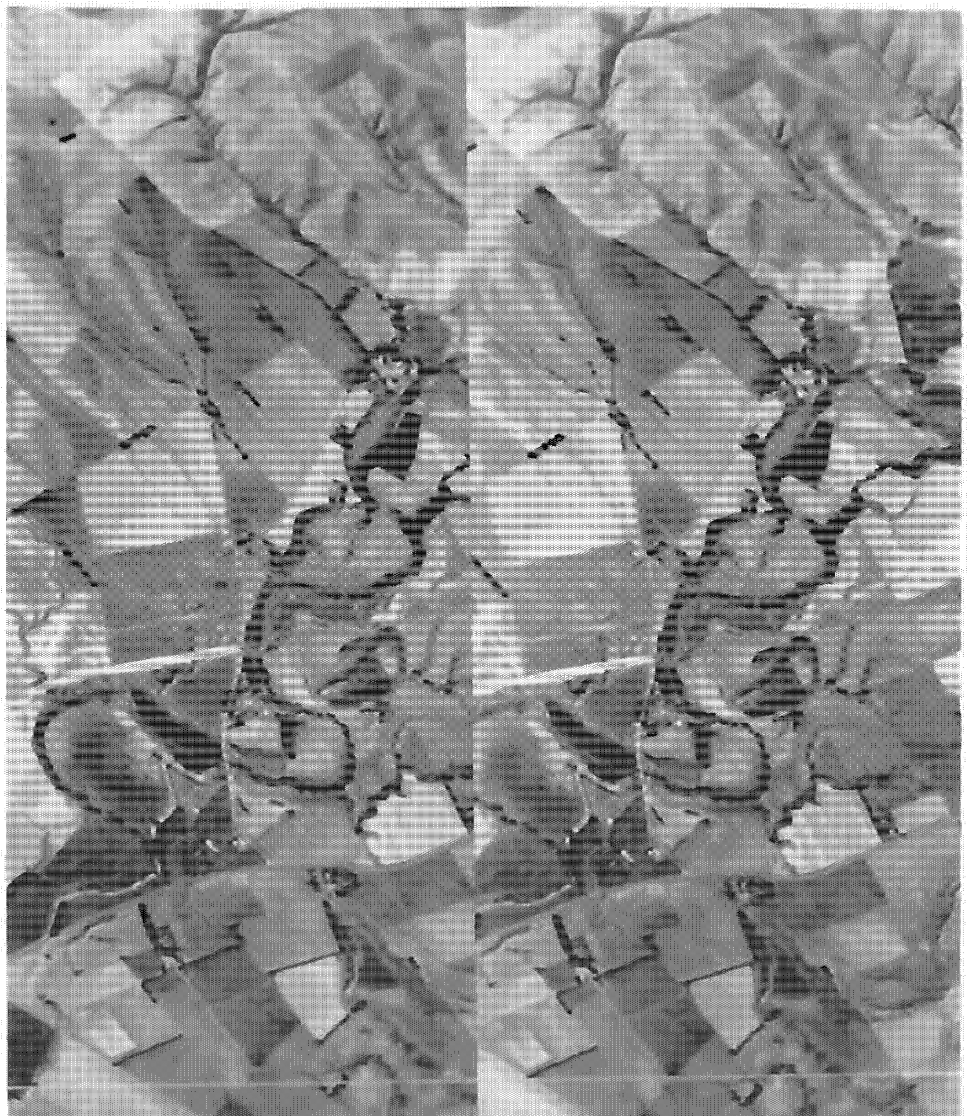
4.3.3 CENTRAL LOWLAND

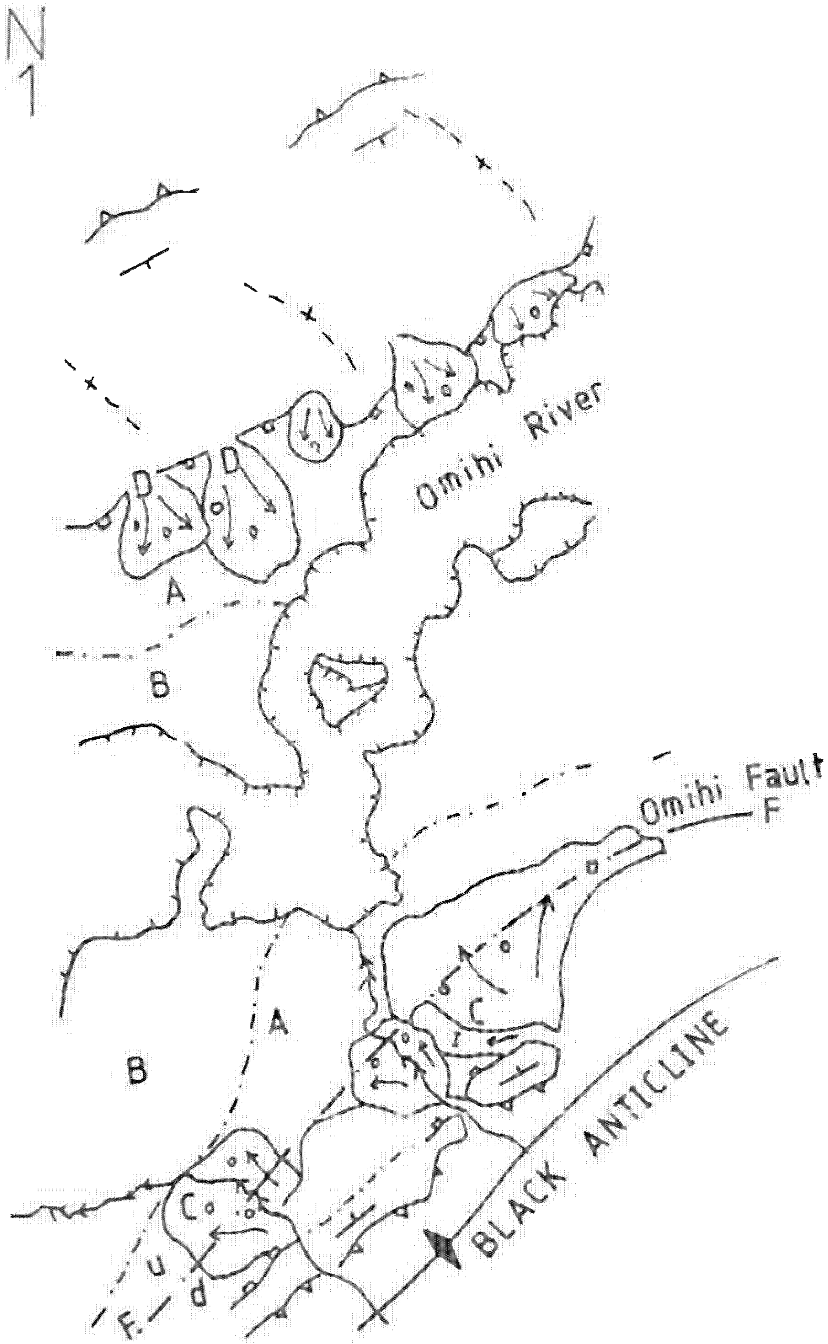
The central lowland has undergone only minor dissection (Stereomodel 12). Waipara River and Weka and Omihi Streams and their tributaries are downcutting into this surface as are several small gullies. The gullies have the distinctive V-shape associated with youth. The present channel of the Omihi Stream and its creek are frequently meandering in character, with a narrow flood plain and have at least five terraces. The flood plain of the Omihi Stream shows clear evidence of contemporary erosion and deposition. Substantial migration of the stream channel has taken place, and meander cut-offs are at a higher level with some small ox-bow lakes preserved (Stereomodel 4). This may indicate recent uplift in the area but may also be a consequence of climatic change. The steep and eroding cliff sections along the river indicate that lateral movement of the channels is occurring. River incision has formed very steep scarps giving the fan surface its elevated position in the landscape of the Omihi valley. This surface is classified into two main levels, and two fan-types may be recognized (active and inactive alluvial fans).

The present channel of the Waipara River is frequently straight and braiding with a broad flood plain zone throughout the Canterbury Surface. A sequence of ten erosional terraces has been identified below the level of the Canterbury Surface, which are discussed in Chapter Six. Weka Stream has five erosional terraces. The present channel of the Weka Stream is incised into the Canterbury Surface with a braided pattern.

Stereomodel 12

Stereomodel of landform in the central lowland area. Note the broad aggraded valley-plain (area B) and incised meanders cut in weak alluvial valley filling. Two fan-types are recognisable in area A.




N
Omihi River
Omihi Fault
F
D
D
A
B
A
B
C
C
u
d
F.
BLACK ANTICLINE

4.3.4 NORTHWESTERN HIGHLAND

On the northwestern side (Fig. 4.4), the highland gradually descends to the floor of the central lowland Canterbury Surface, with no sharp boundary existing between them (Stereomodel 13). Most of this terrain is outside the study area. Typical asymmetric homoclinal ridges are well illustrated in the geomorphic map (Plate 2). Slopes of 15^{o} - 30^{o} and parallel dendritic patterns are characteristic of this terrain.

4.4 MASS MOVEMENT DEPOSITS THROUGHOUT THE AREA

4.4.1 FACTORS CONTROLLING SLOPE FAILURE

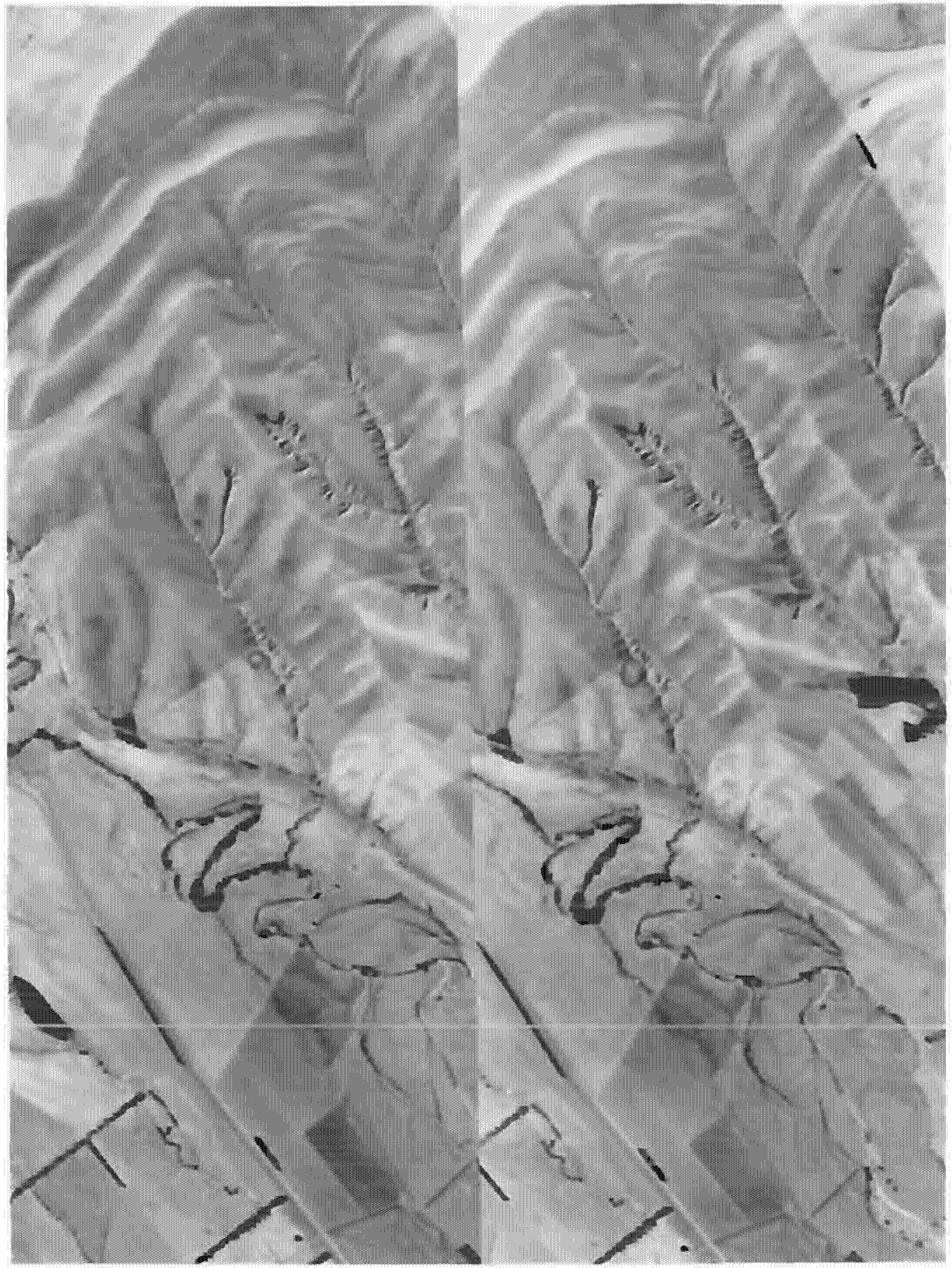
The following discussion is mostly concerned with processes that are associated with mass movements taking place on steep slopes ($>15^{o}$).

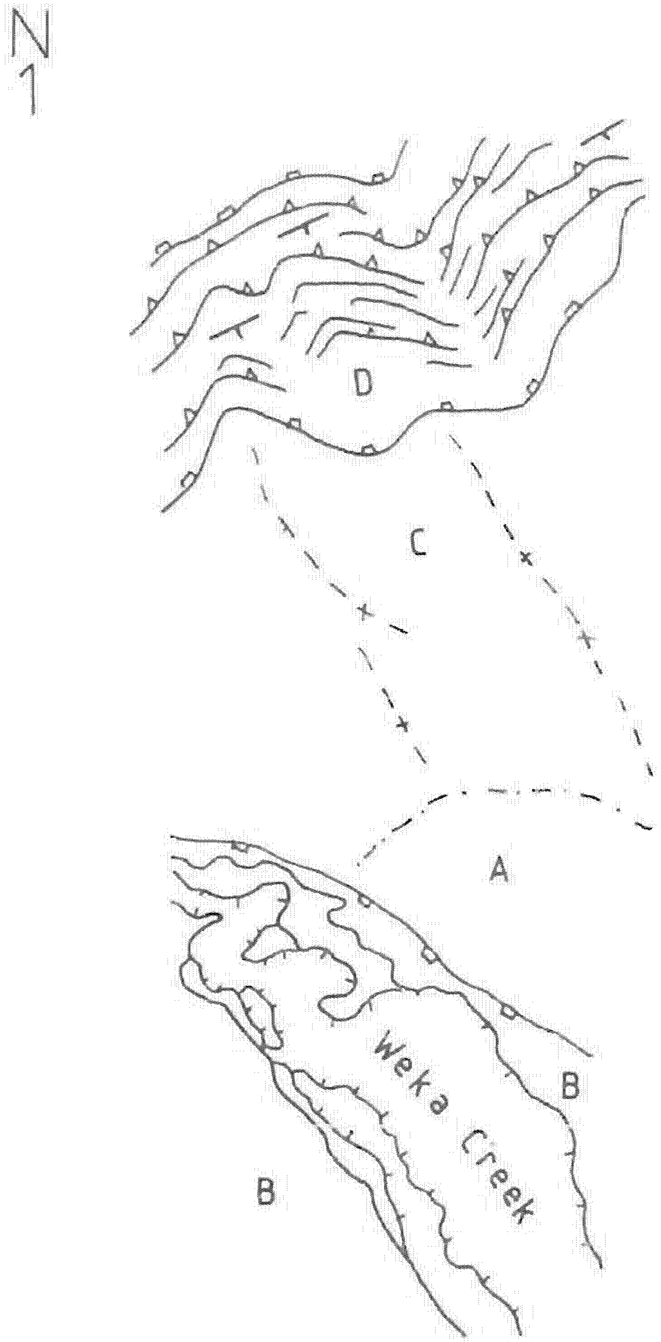
One of the most striking correlations observed was that between mass movements, degree of slopes, lithology and the principal compression direction of the neotectonic stress field in the study area. Amuri Limestone, Ashley mudstone and Greenwood Formation etc, are formations that are most highly susceptible to landslides. Numerous slides are evident throughout the study area, some due to undercutting by streams and others to steep slopes or stress release along joints and faults.

Mass movements occur largely in three of the 15 geomorphological units discussed before, namely the homoclinal structures, backslopes and coastal cliffs. Thus, although slope angle is an important variable in the analysis of the distribution of mass movements and their causative factors, photo study allows for the recognition of different levels of erosional activity for slopes of similar gradient

Stereomodel 13

Stereomodel representing the transition zone between the north-western highland (Weka Pass) and the central lowland (Canterbury Surface). The morphology of the crest, slopes, rock types, and gulley forms are quite characteristic for each lithological formation. At D, hogback ridges are formed in resistant Greta Formation whereas lithological boundary is not sharp between unit C and A.




N
D
C
A
Weka Creek
B
B

in various parts of the terrain. Likely modes of failure in relation to trigger mechanisms, such as landslips, rockfall or the progressive weathering of oversteepened slopes, can be identified. The shallow slides occur in surficial material which has lost its cohesion as a result of weathering and consequently the slope maintains a stable angle appropriate to the angle of shearing resistance of the material. These slides are more common along the dip slope or recently incised valleys. These can often be distinguished by careful aerial photographic interpretation and field observations. Types and distribution of actual processes and their geomorphic traces are indicated by line symbols on the geomorphological map (Plate 2).

4.4.2 DESCRIPTION OF NOTABLE EXAMPLES

a. Landslides

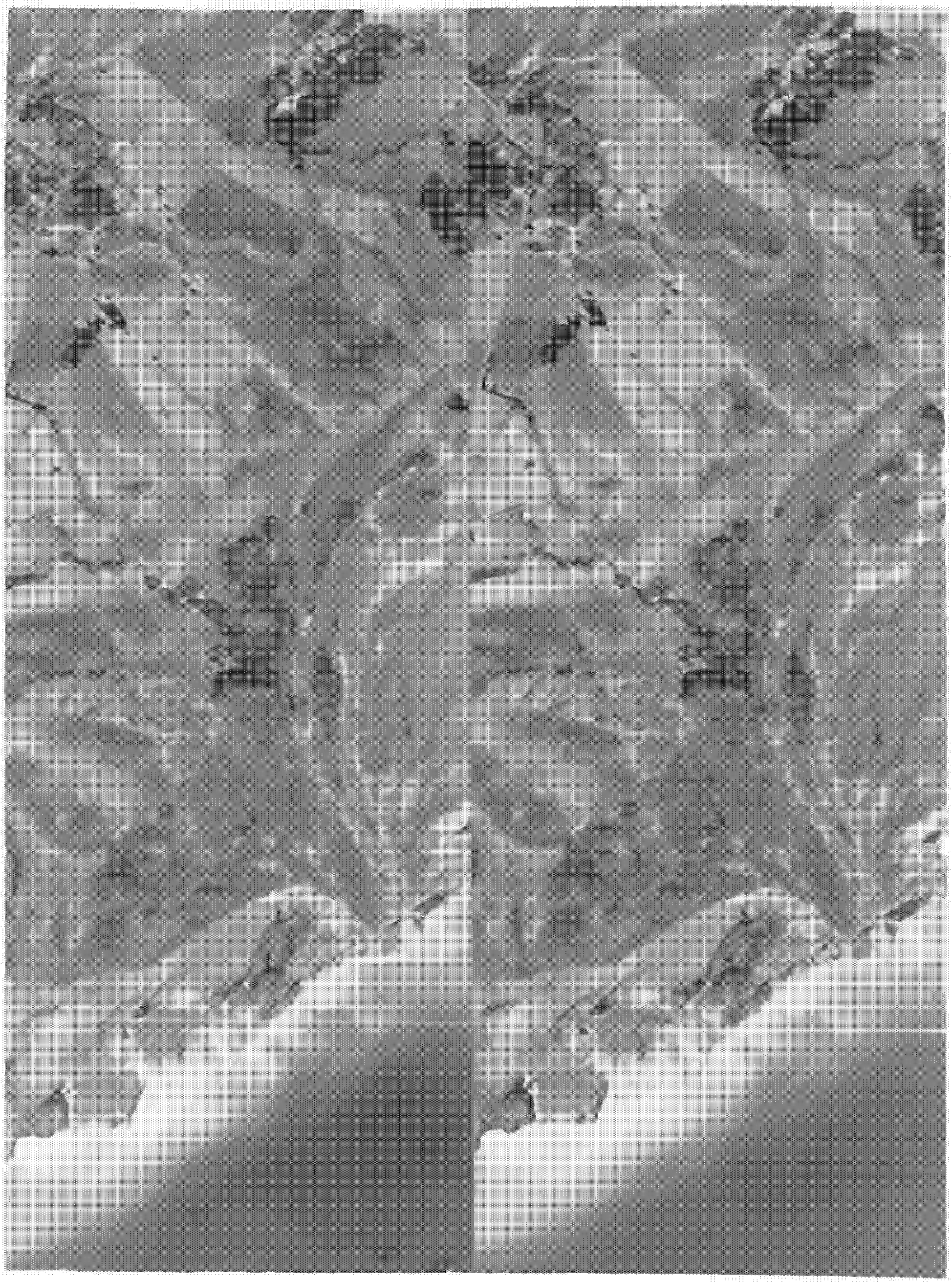
i. Montserrat Slide

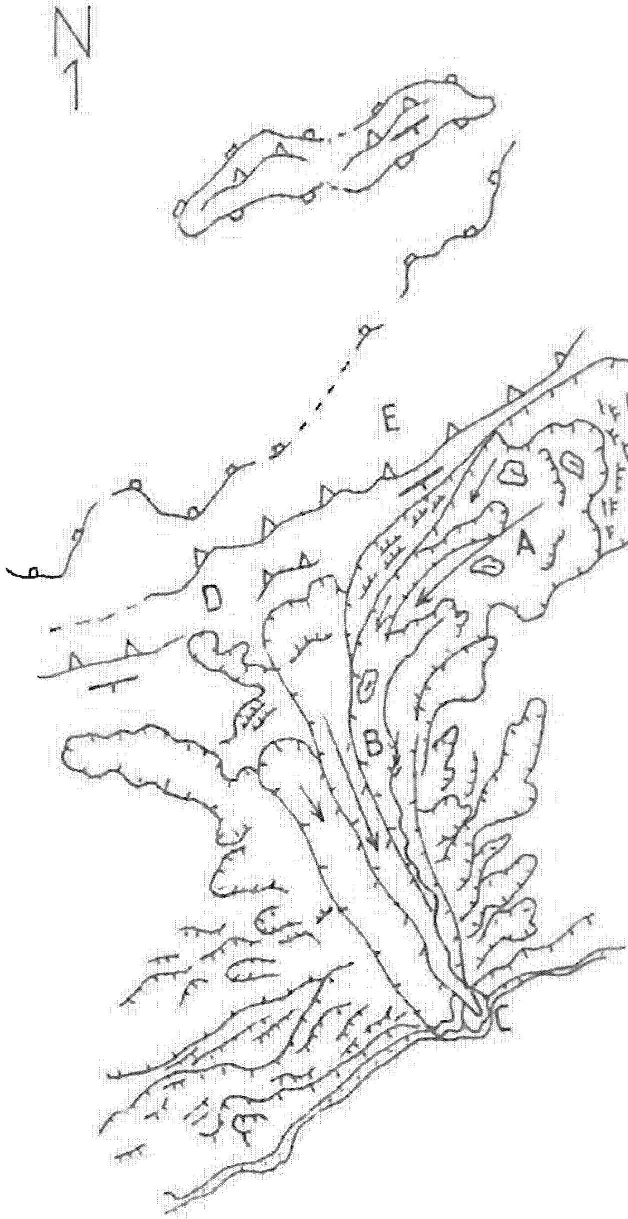
The largest landslide is that on the south eastern limb of the Montserrat Anticline (Stereomodel 14) and is situated in the Tertiary cover rocks consisting of siltstones, mudstones, sandstones, and limestones. The failure appears to be deep seated, originating in the bentonitic mudstones of the Ashley Mudstone and Amuri Limestone (Morris 1983). The Montserrat landslide is a massive complex class VI slump-earth flow (Varnes 1978 classification), but rock slumps, topples, debris slides, and earth and rock block slides occur throughout the area on a much smaller scale.

Morris (1983) made a detailed investigation of the Montserrat landslide. He reported that the headscarp region appears to be primarily a deep seated failure, whereas the lower half of the slide seems to be debris flow on a continuous slide plane. The flow debris

Stereomodel 14

Montserrat Landslide. A characteristic expression of a massive complex class VI slump-earth flow. Note the change in the direction of the flow axis (area A and B). Cliffs above the toe (area C), at the southern end of the landslide show offset beds which are the result of either a massive rotational slumped block, or a fault.





N
E
D
A
B
C

is characterised by hummocky disturbed ground made up primarily of limestone and mudstone colluvium. The abundant seepage areas and swamps are the result of poor drainage and help proliferate the continued slumping.

The mean slide direction has changed with time. From the head scarp to the center, the direction is east-northeast; then from the center to the toe, its direction changes to a north-south bearing. The reason for this may be that the head scarp has migrated following the bedding plane of the Amuri Limestone ridge. Now ancient areas of slope movement can be seen on aerial photographs which are well covered in vegetation and have degraded into an undulating slope.

ii. Inland Motunau Slide

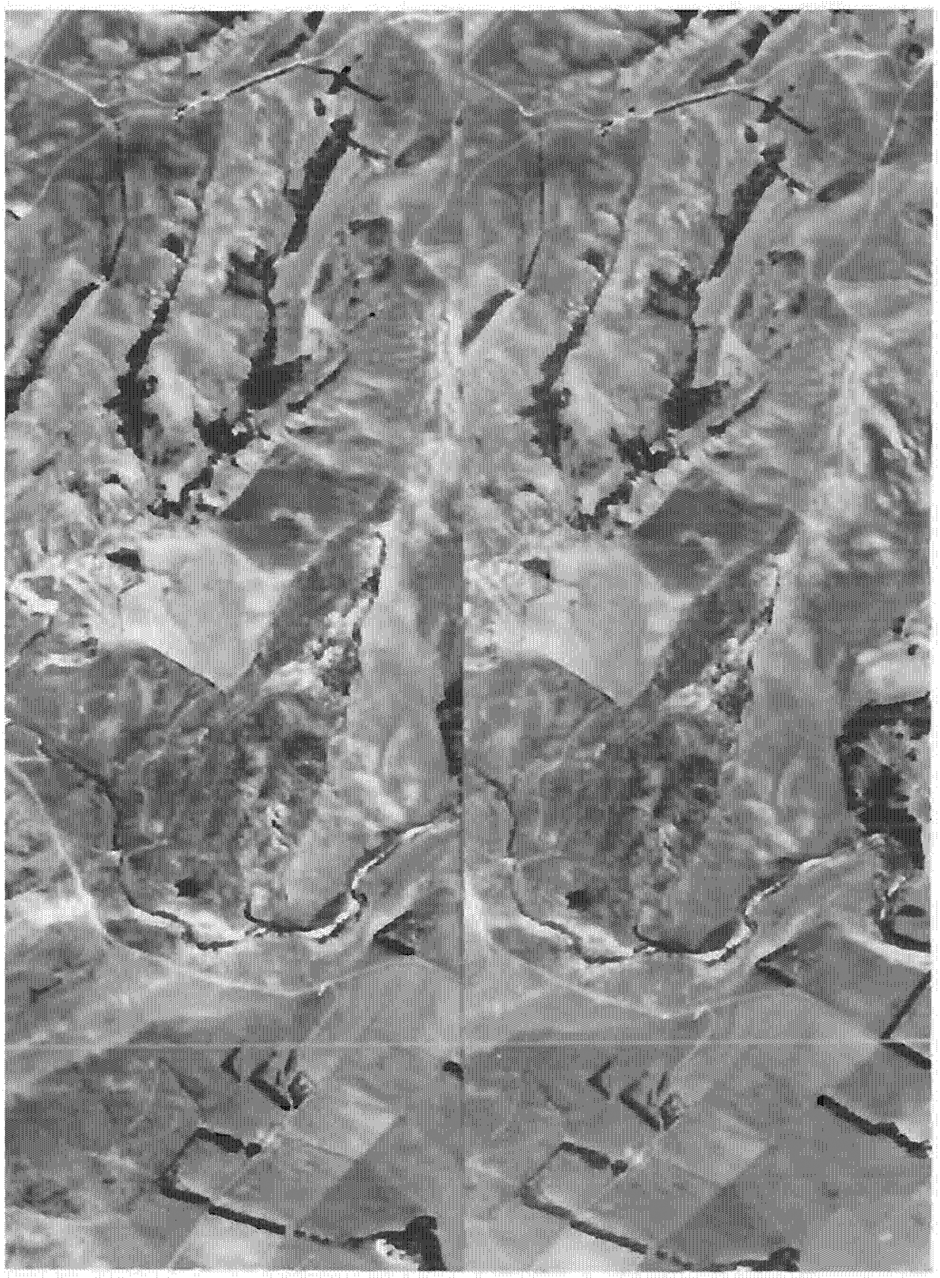
A landslide similar to, but smaller than, the Montserrat slide occurred 3.5 km inland from the Motunau beach. Although the same geologic controls (bedding, structure, and lithology) allowed the Montserrat and this landslide to occur (Stereomodel 15), there is a fundamental difference in their mode of occurrence. The movement vector for the Montserrat landslide is at right angles to the regionally uniform strike of the strata, that is NE-SW except at the head scarp region as I mentioned before. In the other example, slide movement was parallel to the strike of bedding. Both types are common in the study area.

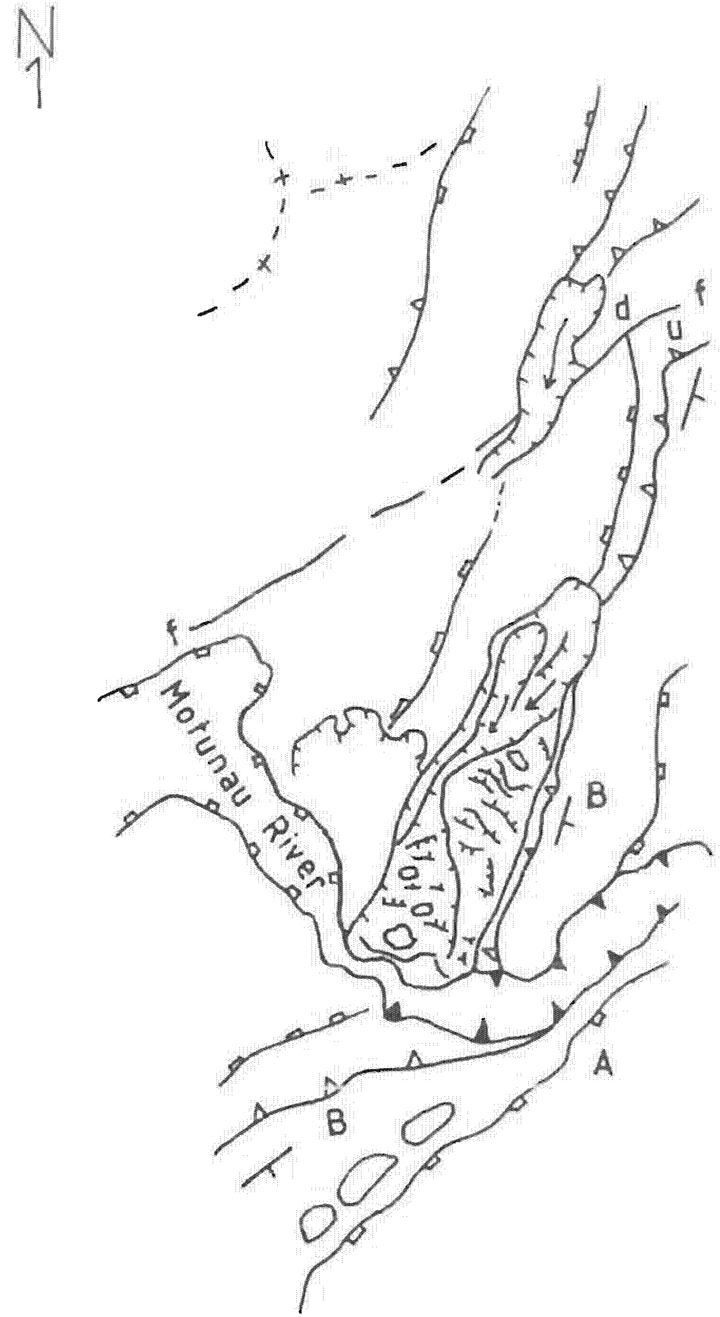
iii. Coastal cliff failures

The failure at the coastal cliff east of Dovedale River (Stereomodel 16) is a typical translational block slide, with the slide surface being a siltstone bedding-plane surface of Mt Brown

Stereomodel 15

Inland Motunau Slide. A characteristic expression of a massive complex class VI slump-earth flow. Slide movement was parallel to the strike of bedding(area B). Note the offset of the hogback ridge by the oblique-slip fault to the north of the photo.




N
f
d
u
f
Motunau River
B
A
B

Formation. The uppermost edge (landward) of the main slide block shows pronounced tension crack development and formation of small horst and graben structures. These were probably developing prior to failure by bedding plane creep, and opened by tensional fracture during the main failure. They are slightly tilted. However the seaward edge of the block is considerably more deformed, although it too remained entire until coastal erosion forces overcame much of it. The highly regular strike of the beds is responsible for the straight line of the main scarp. It has become modified by secondary failures and gully deveopment. Translational block slides are common along the coastal cliff. Their distribution is illustrated in Plate 2. Smale et al. (1982) reported a bedding-plane landslide near Mt Vulcan where a block of massive sandstone, 4 hectares in area and approximately 15 metres thick, slid down.

b. Rock falls

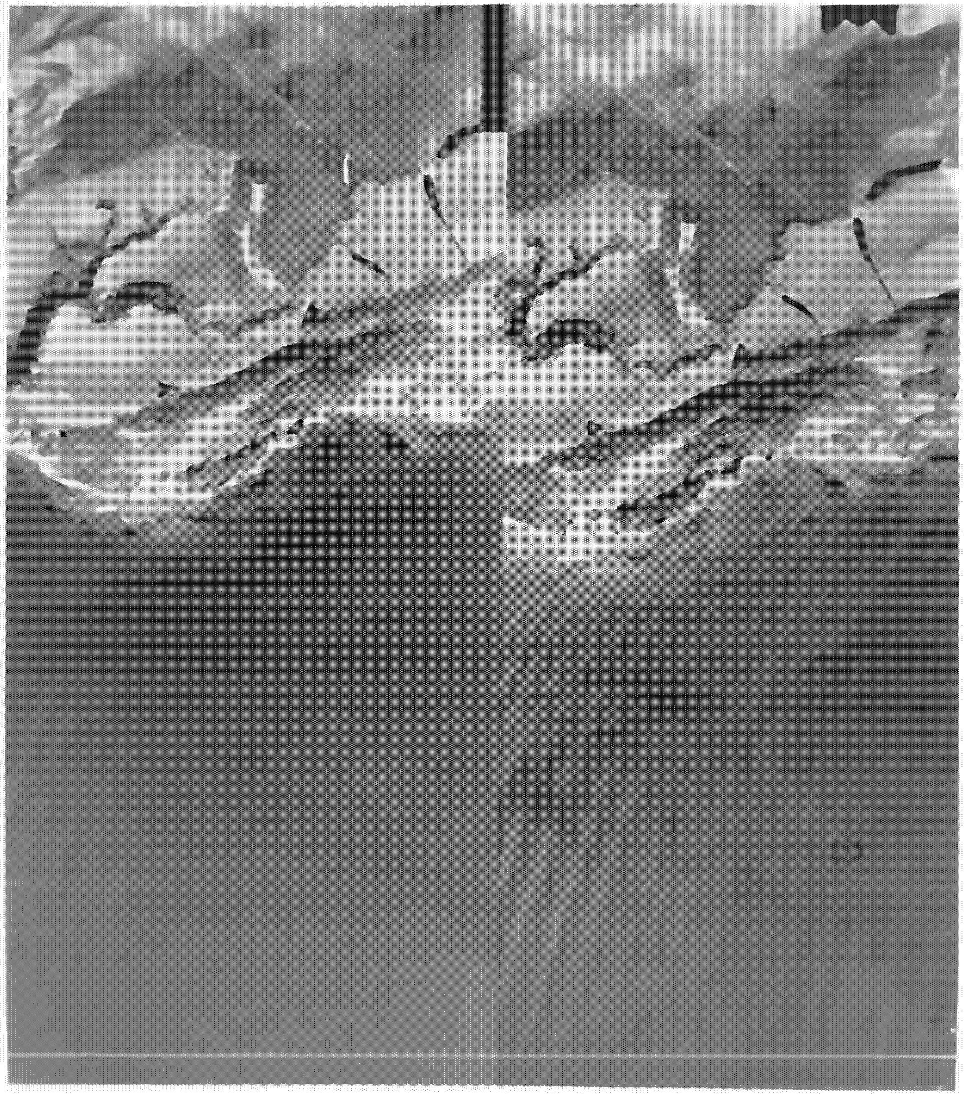
Rock falls from the steep cliffs of the Amuri Limestone, Omihi Formation and Greenwood Formation escarpments were seen. They include recently fallen, highly angular blocks not yet subjected to prolonged weathering.

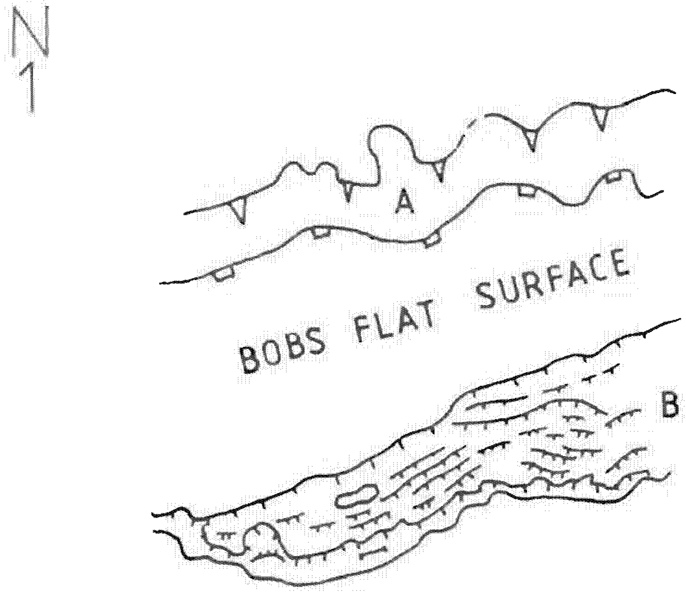
4.5 SINKHOLES

The development of sinkholes in limestone has often been ascribed to solutional activity (Jennings 1971, Sweeting 1972), and are commonly observed on the Omihi Formation. The best of these features are shown on Plate 2, along the northeastern limb of Cass Anticline and around the plunge zone of Montserrat Anticline (Stereomodel 1&2).

Stereomodel 16

Typical example of translational block slide at the coastal cliff east of Dovedale River (area B). Principal motion was by sliding along gently seaward dipping strata. Note the raised, dissected, marine terraces (Bob's Flat Surface) and former sea cliff (area A).




N
A
BOBS FLAT SURFACE
B

CHAPTER FIVE

MORPHOMETRIC ANALYSIS

5.1 INTRODUCTION

Topography is the most extensive and readily available record of recent vertical crustal deformation at the regional scale. As the need for information on recent tectonics grows so does the need for methods of topographic analysis. Hence, many investigators have utilized morphometric parameters in order to establish relative and in some cases absolute uplift/tilt rates and ages (Volkov et al. 1967, Stearns 1967, Bull & McFadden 1977, Wallace 1977a & 1978, Bucknam & Anderson 1979, Mayer 1979 & 1984, Colman 1983, Seeber & Gornitz 1983, Nash 1984, Hanks et al. 1984 & 1985, Sterr 1985, Crittenden & Muhs 1986.

The longitudinal valley floor profile is the least transient expression of fluvial processes, reflecting as it does, geological influences such as the effect of vertical tectonic displacements of valley floor and base-level change, of climatic change on the processes of erosion and deposition, of the distribution of outcrops of different lithologies, discontinuous changes in sediment calibre downstream, pool and riffle sequences, and to the increases in discharge due to entry of large tributaries. Fundamentally, the longitudinal profile of the river has an overall concavity that reflects a progressive decrease of gradient, so that a concavity-convexity index describing profile form, locates segments affected by some of these modifying influences.

There are ample reasons to pursue the rational description and prediction of river profiles. If we can adequately describe the ideal, graded forms for particular rivers, then deviations from those forms can be considered as signifant sources of information, Snow & Slingerland (1987).

In a detailed analysis, Leopold & Maddock (1953) explained the concavity of the longitudinal profile in terms of the downstream decrease of sediment load in relation to discharge. Hack (1974) concluded that anomalously high gradient index values may occur on a reach of high energy which could correspond to a belt of resistant rock, high roughness, tectonic warping or to any erosional disequilibrium either where increased competence is needed to move coarser material or where non-uniform discharge increase in the downstream direction.

The study on which this section is based had two related objectives. The first was to determine the relationship of altitude and gradient index to lithology and structure in two selected areas, so as to assess the extent to which erodibility of rock has influenced the morphology of the whole study area. The second objective was to determine the relative rate of uplift of marine terraces from place to place. Both ojectives which, as will be shown, are related were attained to a first approximation. Within the areas covered in these analyses, it is apparent that altitude and stream pattern are related closely to the lithology and structure. The degree of adjustment of drainage to underlying rocks is substantial.

5.2 EQUIPMENT REQUIREMENTS FOR THE CURRENT RESEARCH

Fundamental to the analysis of stream gradients was the retrieval and manipulation of large amounts of altitude and location data. The speed of this process was greatly facilitated by use of a digitizer interfaced to a minicomputer which allowed all stream courses to be scanned rapidly and the co-ordinates of all critical points along each stream to be stored and retrieved for computer based computational procedures.

Several pieces of equipment greatly reduce the time and effort needed to extract data from maps and photos. The Department of Forestry Work Station facility used for this study comprises the following equipment:

5.2.1 DIGITIZER TABLET - this piece of equipment is an electronic scaling device, similar in appearance to a draughting board (see Fig. 5.1). The function of the digitizer is electronically to scale co-ordinates off documents taped to its surface and then transmit these coordinates to the mini computer. The digitizer provides a very accurate (+/- 0.1 mm) and rapid (up to 30 coordinates per second) method for extracting data from maps or photos. Make and model of the digitizer tablet is NUMONICS 2300, U.S.A.

5.2.2 GRAPHICS PLOTTER - this piece of equipment (see Fig. 5.1) is used to plot extracted map data in a form which is useful to the planner. The plotter greatly reduces the time needed for plotting data and the likelihood of error. Make and model of the plotter is IWATSU SR 6602 X-Y Multipen Plotter, JAPAN.

5.2.3 GRAPHICS TERMINAL - this piece of equipment, shown in Fig. 5.1, is used for two purposes:

1. to type in additional data through the keyboard, and;
2. temporarily to display graphical data. The graphics screen

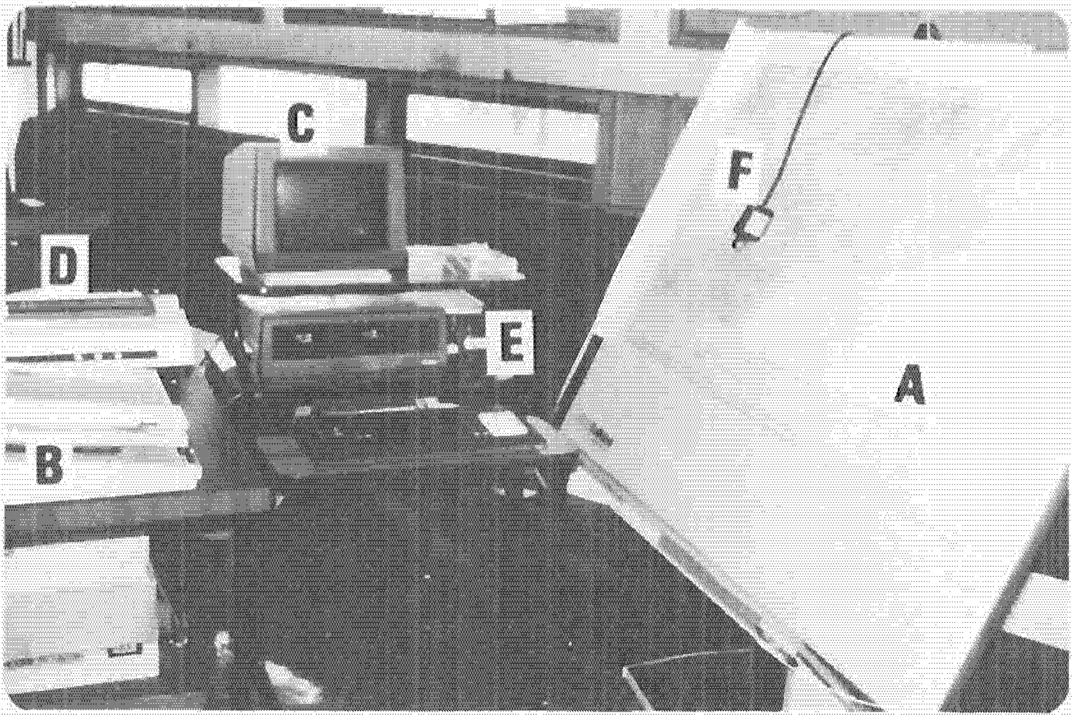


Figure 5.1

The digitizer device

Department of Forestry Work Station, University of Canterbury.

It has the following components:

A. Digitizer Tablet, B. Graphics Plotter, C. Graphics Terminal, D, Printer, E. Minicomputer, F. Tracing Device.

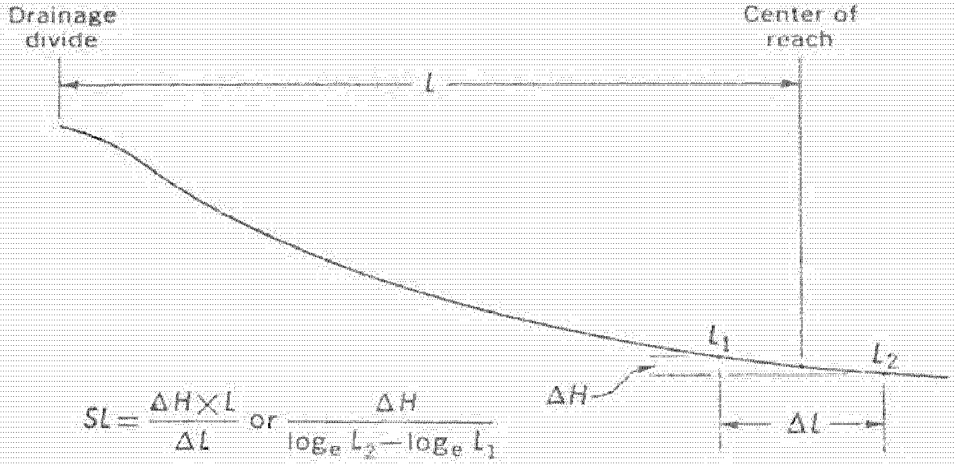


Figure 5.2

Measured parameters used in calculation of gradient index. Symbols are defined in text.

is often used to examine data before plotting it on the graphic plotter.

5.2.4 PRINTER - this piece of equipment, shown in Fig. 5.1, is used to print tabular results. Make and model of the printer is C-ITOH 1550 DOT MATRIX, JAPAN.

5.2.5 MINICOMPUTER - this piece of equipment - shown in Fig. 5.1, drives all the other pieces of equipment together as a working unit. The computer performs calculations on the data supplied by the digitizer or terminal and then sends results to either the graphics screen or for printing. The computer is self-contained and requires no connection to a remote computing center. Make and model of the computer is SIRIUS 1, U.S.A.

5.3 METHODS OF STUDY AND DEFINITIONS OF FACTORS MEASURED

Inasmuch as expressions for measurable elements of a river system are used throughout the report, it is desirable that they be explained at the outset. Measurements made at more than 3000 localities constitute the data for analysis of the factors controlling stream profiles. The measurements described below relate to a single locality, or to a point on a stream channel.

5.3.1 CATCHMENT

The term catchment refers to the drainage basin of the principal stream and of all the tributaries which enter it.

In this section, two catchments (16 and 27) were selected as a sample study. In practice, the catchment area can be measured on topographic maps or on aerial photographs, by using a digitizer device, but this measurement is currently under development.

5.3.2 LENGTH

The term length, denotes the distance from a locality on a stream to the drainage divide at the head of the longest stream above it. The distance can be measured on the map in any unit desired.

5.3.3 THRESHOLD OF CRITICAL POWER

A geomorphic threshold is a transition point or period of time that separates different modes of operation within part of a landscape system (Bull 1979). For example, degradation and aggradation for streams are separated by the threshold of critical power (Bull 1979), defined as a ratio equal to one where the numerator consists of those variables that if increased favour degradation (stream power), and the denominator consists of those variables that if increased favour aggradation (resisting power) as shown in Fig. 5.3a. Both climatic and tectonic change can affect the components of the threshold of critical power (Bull & Knuepfer 1987). Uplift tends to increase slopes, and cooler and/or wetter climates generally increase stream discharge. Changes in the amount and size of sediment yielded from hillslopes may be caused by either tectonic or climatic changes, but resisting power also varies with changes in hydraulic roughness that are independent of tectonic and climatic controls (Bull & Knuepfer 1987).

5.4.4 GRADIENT INDEX

The gradient index can be measured on topographic maps, or on aerial photographs using photogrammetric methods. This index provides

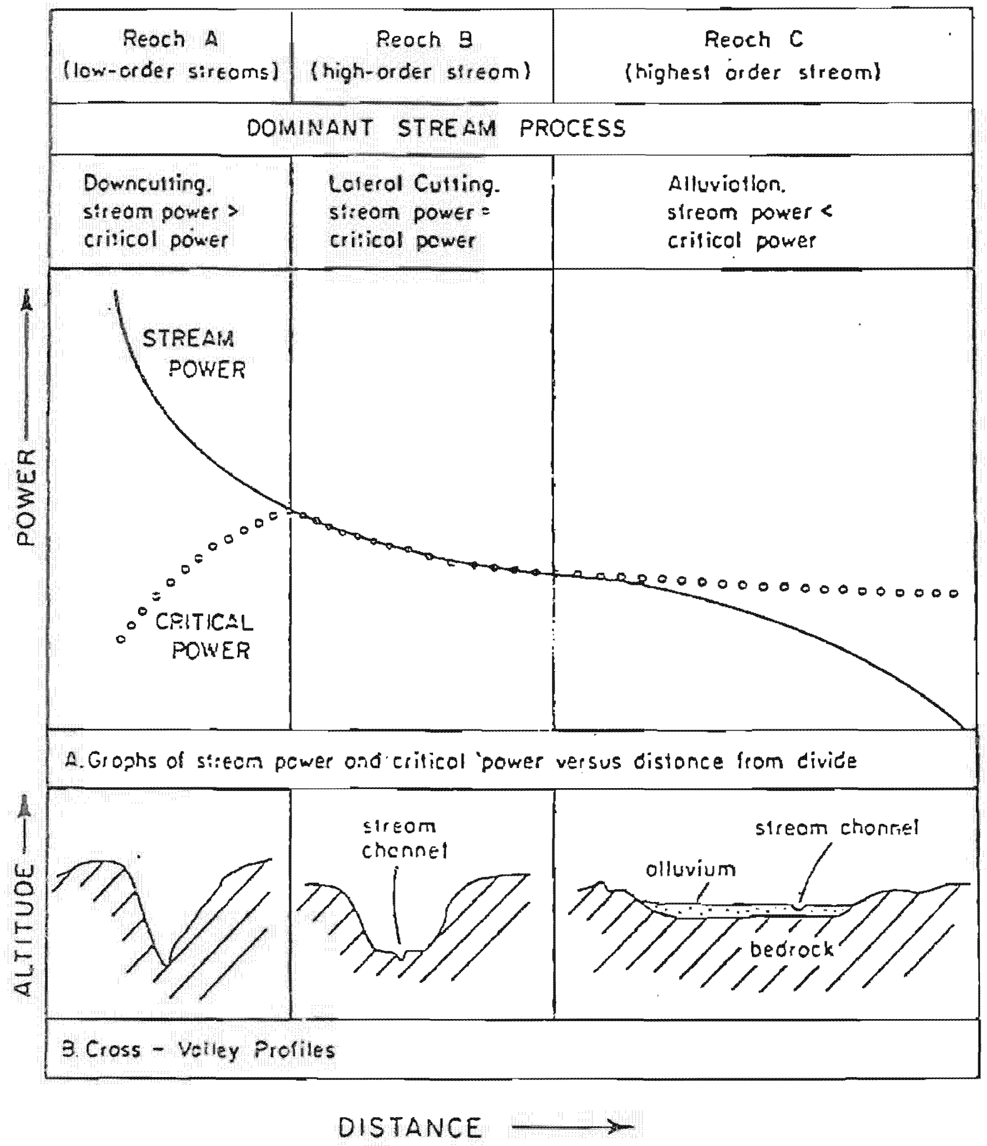


Figure 5.3 Diagrammatic sketches and graphs of stream power and critical power for drainage basin (from Bull 1979).

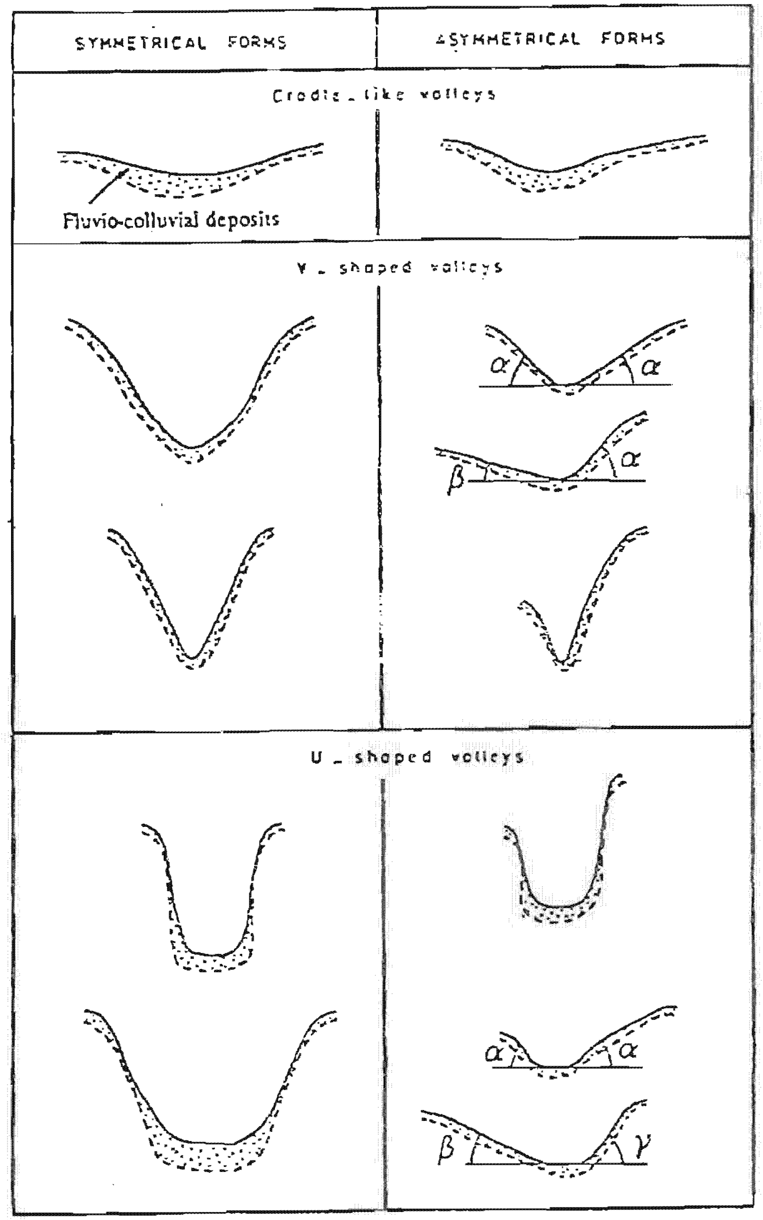


Figure 5.4
Various cross-valley forms.

a basis for comparing stream reaches of different sizes and is believed to reflect stream power. The gradient index is proportional to the slope of the plotted profile. Its numerical value can be calculated from the tabulated data at any point (see Appendix 1).

The formula for the gradient index (GI) is:

$$GI = \frac{\Delta H \times L}{\Delta L} \quad \text{or} \quad \frac{\Delta H}{\text{Log}_e L_2 - \text{Log}_e L_1} \quad \ldots\ldots (1)$$

The parameters measured are shown in Figure 5.2, where ΔH is the difference in elevation between the ends of a reach, and ΔL is the length of the reach, and L is the length measured from the drainage divide at the source of the longest stream in the drainage basin above a locality on a reach (Hack 1973a).

The gradient index of a specific reach can be evaluated by comparison with the gradient index for the whole river profile defined by:

$K = Hi - Hj/Ln\ Lj - Ln\ Li$.....(2)

where i and j refer to two points at the two ends of the river profile, and K is the gradient index of the entire profile (Seeber & Gornitz 1983).

In this study the gradient index numbers are calculated from (I) for each reach between the points at which the stream crosses consecutive contours. The gradient index number of each such reach is compared to the gradient index (K) of the entire profile calculated from (2) see Appendix 1. The entire profile is defined as lying between where Hj and Lj are taken at the downstream limit of the river (the mouth, or if a tributary, at its junction with the next lowest stream) and Hi and Li are taken the highest topographic contour crossed by the river. In catchments 16 and 27 I have distinguished reaches where the gradients are significantly steeper (GI/K>2) and much steeper (GI/K>4) than the ideal profiles (Appendix 1). The profiles are also subdivided into reaches where the gradient index varies relatively little (GI/K<2). The significance of anomalous gradients can be evaluated throughout the study area.

5.3.5 KNICKPOINT

The knickpoint as it will be referred to here, is a location on the longitudinal profile, where there is an abrupt change of elevation and slope. Each channel reach that contains the knickpoint has a high (convex) or low (concave) gradient index value. This feature migrates up the valley floor or along an existing channel with time.

5.3.6 LONGITUDINAL PROFILE

The longitudinal profiles of many graded rivers have smooth, concave-upward curves similar to graphic curves produced by a variety of simple mathematical functions. In various studies mathematical curves of exponential, logarithmic, and power function forms have each been fitted to actual river profiles with success (Woodford 1951, Hack 1957, Butakov 1970, Shepherd 1985), but none of these equations has been shown to be universally applicable in the field. Such variation in field results ought to be explicable in a more comprehensive analysis of stream longitudinal profile form than has yet been attempted.

Long-profile characteristics are strongly scale dependent. Consecutive reach slopes combine to create the complete long profile which reflects long-term geological development influenced by tectonic, lithologic and morphogenetic histories as well as the recent channel pattern adjustments (Appendix 1).

Individual profiles can be studied graphically by simple plotting on graph paper. The vertical co-ordinate is an arithmetic scale and represents the elevation above sea level. The horizontal coordinate could be represented as a logarithmic scale and represents the stream length or distance from the source, which will have the effect of representing an exponential curve as a straight line as proposed by Hack (1937a).

River profile data are extracted from the topographic map (Plate 3) using the digitizer and displayed on the graphics terminal or plotter. In practice it was found to be more useful to plot the horizontal coordinate in arithmetic format as developed for the computer programme (see Appendix 1) and allows more ready comparison between rivers of different sizes by adjusting the scale of the individual rivers such that they occupy the same distance along this axis.

The digitizer records the coordinates of elevation points by running down the stream course on the map with a tracing device and recording the contour elevations and their distances from the head of the stream at appropriate points. It is important that the distance (L) be measured along the principal (or longest) stream above the locality. The operator must always proceed downstream from the head and never continue past a stream junction with a longer stream than the one being measured.

Many of the opportunities for error associated with the use of maps, such as misreading elevations or coordinates are eliminated through the use of the digitizer and the operation is much faster than manual methods.

These data could of course be presented and analysed in various different ways, but here representation is confined to the simple plotting of height against distance to construct linear surface profiles of the various rivers.

Unless dominated by structural or tectonic controls, the longitudinal profiles of numerous rivers measured in this study are, overall, concave upward, but may have straight or even convex-upward segments (see Appendix 1).

Additionally, individual profiles can be studied graphically by simply plotting distance against gradient index. Distance is plotted on a linear scale and gradient index logarithmically. Any changes in the value of the gradient index along the longitudinal profile can be accentuated more accurately by this plotting. This is well illustrated in Appendix 1.

Reaches where the stream power has either increased or declined can be readily identified in the profiles. These segments can be interpreted in terms of transitions across the threshold of critical power and with changing of the river cross-section.

5.3.7 CROSS SECTIONS

A change of channel cross-section is indicative of the power of a stream to maintain its channel in response to lithology and to changing base level, thus a study of valley cross-sections provides evidence for periods of rejuvenation. Three fundamental profiles separate critical threshold conditions (Fig. 5.4).

In practice, valley and channel cross-sections have been sketched from stereoscopic airphoto study. Accordingly, vertical scales of the cross-section profiles tend to be considerably exaggerated, but the three basic cross-section types can still be distinguished. They are:

i) Smooth, cradle shaped:- produced by low erosional force where the rate of accumulation is greater than the rate of degradation

ii) sharp V-shaped:- produced by strong, vertical erosion which may occur in recently uplifted areas, or may be caused by more resistant rock types.

iii) smooth or sharp U-shaped:- produced probably by a period of strong vertical erosion followed by a rapid slowing or pause in the rate of downcutting. Usually the U-shaped valley is partly infilled with sediment (excluding glacial processes).

Any increase in the river power will rejuvenate the vertical erosion along the valley bottom and will produce a new composite type. This criterion probably indicates a recent uplift in the area but may also be a consequence of climatic change. In short, the evidence supplied by the cross-sections, should correlate with the long profile reconstruction.

5.4 DISCUSSION

The longitudinal profile of a stream is a property of stream geometry that can provide clues as to underlying materials as well as insights into geological processes and the geomorphic history of an area facilitated by gradient index analysis (Hack 1960, p.89). These studies were made on 63 catchments, the streams range from gently to steep sloping terrain.

Bull (1979) demonstrates the widespread application of the threshold approach to the understanding of the interrelations between processes and landforms. He states that, "The critical power threshold separates the modes of erosion and deposition in streams and is dependent on the relative magnitudes of power needed to transport the average sediment load and on the stream power available to transport the load". This approach emphasizes the morphological point of view. In reaches where stream power exceeds critical power, vertical erosion predominates, but lateral erosion predominates where a stream is close to the threshold as shown in Figure 5.3A and B. In practice, two conditions relating to the critical power threshold can be recognized easily in the longitudinal profiles: (1) reaches where the threshold has been exceeded, and (2) reaches that approximate the threshold or where the critical power exceeds the stream power. On the arithmetic graphs the knickpoints are observed in those reaches where there is change in the values of the gradient index accompanied by a change in the river cross-section. These criteria are indicative of the power of a stream to maintain its channel in response to lithology and to changing base level. Thus, active downcutting by the stream and lack of evidence for alluviation are clear evidence that the threshold is being exceeded.

Reaches of high gradients or knickpoints along rivers can originate in several ways. They can develop from a eustatic lowering of the base level of erosion (usually sea level), which steepens the gradient in the newly exposed level of the river and increases the amount of down-cutting there. The resulting high gradients are not stable and will migrate upstream by headward erosion and disappear, as the river is regraded to the lower base level. The deep gorges at the end of the coastal catchments (Plate 2) are partially attributed to the eustatic lowering of the sea level. This steep slope may also be attributed to the differential uplift. In this case the role of differential erosion is secondary to the role of tectonics in shaping the profiles of these rivers because, after substantial uplift, the critical-power threshold may be exceeded along an entire drainage net indicated by lack of net alluviation in narrow V-shaped valleys. Once the rivers are entrenched in deep gorges, they tend to maintain their courses throughout further tectonic evolution. Consequently, the drainage pattern is most sensitive to the early stage of tectonic activity.

Sharp changes in gradient in a longitudinal river profile can also occur where the resistance to erosion of the bedrock changes. In the study area, the correlation between lithologic contacts of erosion-resistant rocks and high river gradient is generally good. But along reaches within the same lithology, knickpoints can be ascribed to tectonics (e.g. faults) where a steep slope can be expected on the upthrown side of the fault.

In each catchment, the rivers under discussion all form connected parts of one drainage system (for example figure 5.6 and 5.13), and in each case the total drainage basin is not large, in physiographic terms. The figured example (Figs 5.6 & 5.13) measuring approximately six kilometers by four in its main axes it is not large enough for hydrological conditions to vary greatly in character over its surface. The catchment areas of the different tributaries of course vary in size somewhat, as will the mean discharge of the various rivers, but this sort of measurement had not yet been attempted on the digitizing equipment.

5.5 CASE STUDIES

5.5.1 DESCRIPTION OF CATCHMENT 16

The area studied is the catchment of the Dovedale River (Fig. 5.5), for more details see Plate 2. The geomorphology of the area consists generally of a moderately inclined structurally controlled hogback (Weka Pass Stone) which descends from 1700 ft in the northwest to about 800 ft at the bottom, and sloping to moderately steep, undulating to rolling terrain (Waikari Formation, Mt Brown Formation and Tokama Siltstone) which descends from 800 ft to 300 ft.

The studies were made on about 37 streams in the area shown on Fig 5.6. In general, the drainage in the Omihi Formation (Weka Pass Stone) is parallel to subparallel dendritic and indicates pronounced slope and structural control. Dendritic drainage is developed on the Mt Brown Formation, Tokama Siltstone and Waikari Formation where rocks offer uniform resistance in tilted rock structure. The many tributaries joining the Dovedale River are of resequent, obsequent and subsequent origin while the Dovedale River itself is probably a consequent. The evolution of tributary development is generally up to third order but in places exhibits only second order, indicating early stages of development.

The longitudinal river profile (Fig. 5.7) of the Dovedale River exhibits knickpoints around an altitude of 1300 ft, 1000 ft, 600 ft and 300 ft. These knickpoints are interpreted as being of three genetic types which typify the main processes responsible for producing breaks in gradient.

a) Local base level change by local deformation e.g. faults.

The first knickpoint at 1300 ft, can be ascribed to faulting as well as the lithologic change from Amuri Limestone to Weka Pass Stone.

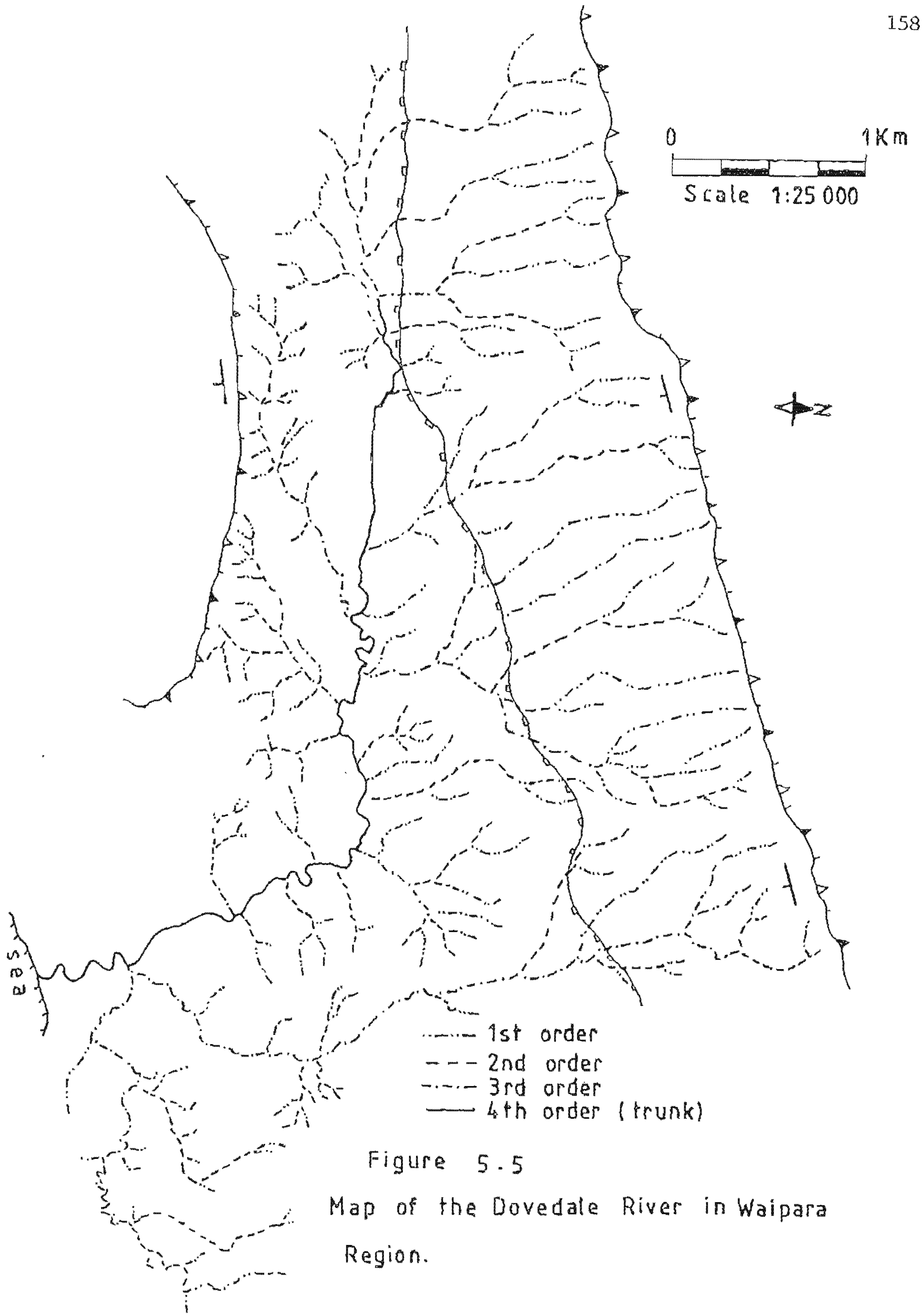


Figure 5.5

Map of the Dovedale River in Waipara Region.

Faults often bring different rock types into juxtaposition so that a hybrid origin for a break in slope is common, as in this case. An interesting feature of this knickpoint is that the critical-power threshold was exceeded only after substantial uplift had already occurred as indicated by a sudden change in the river cross-section (see Figure 5.8 B,C). A sharp V-shaped cross-section is indicative of the power of a stream to maintain its channel in response to lithology, and a steep slope can be expected on the upthrown side of the fault. Amuri Limestone exhibits a low resistance to erosion so that the resulting valleys are narrow at the lower ends and widen upstream into the upthrown block. Stereoscopic study shows this feature accurately (see stereomodel 8).

b) Lithologic control

The second and third knickpoints 1000 ft and 600 ft respectively, coincide with lithological variation (Fig. 5.7). Analysis of the gradient index shows that differential erosion of rocks of different resistance is the major factor that control the development of today's long profile forms.

c) Regional base level change (eustatic and/or climatic and/or tectonic)

For the south eastern margin of the study area there is a fairly direct relation between the differential uplift and the main physiographic features. There are marine terraces superimposed on the Mt Brown Formation and Tokama Siltstone (Plate 2). They have been studied in detail by Carr (1970). Today streams flowing across the southern margin of the area are entrenched into the old marine surfaces. Here the Dovedale River has downcut over 190 ft (60 m) into Mt Brown Formation (see Fig. 5.8 K,L). It will be noted that no

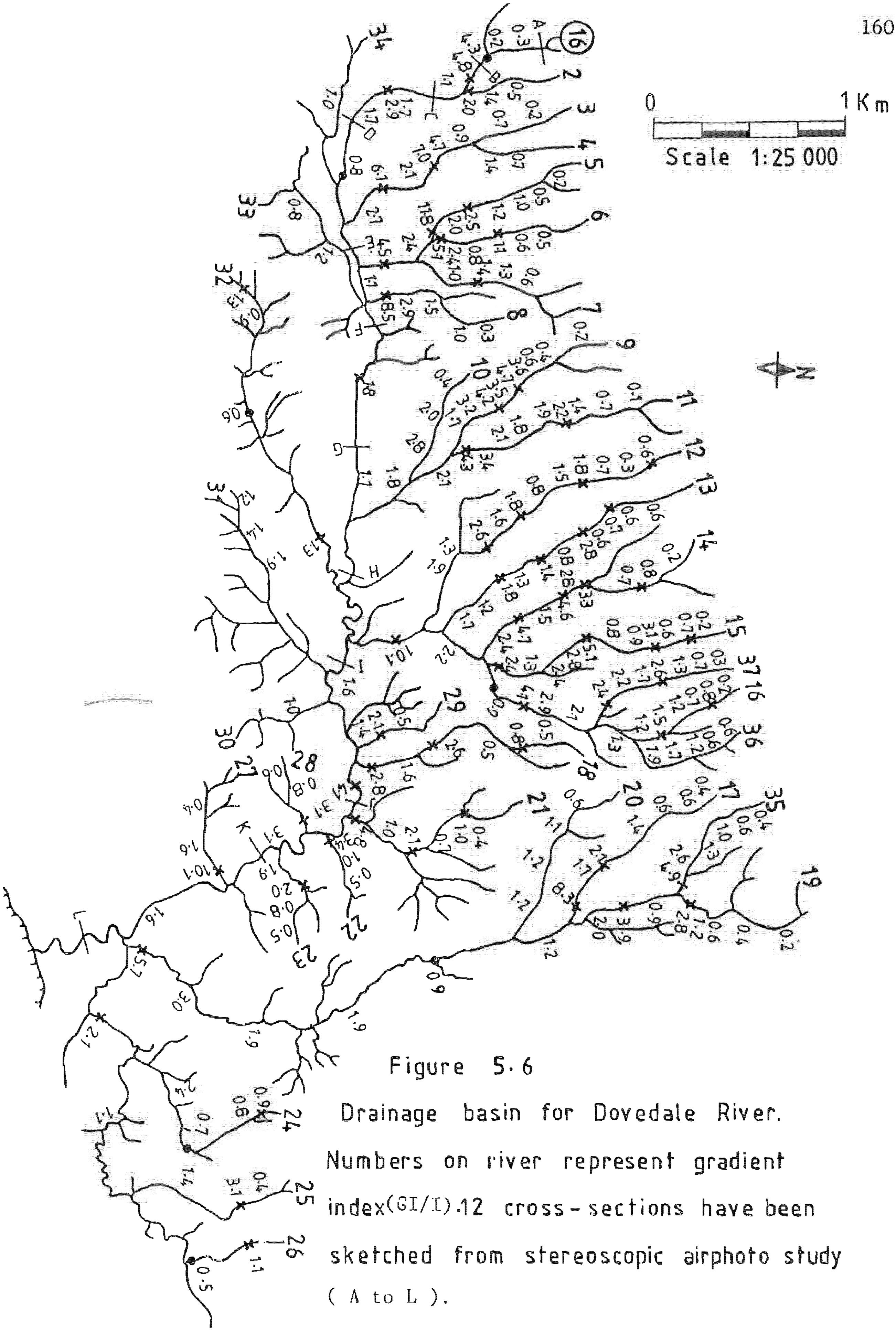


Figure 5.6

Drainage basin for Dovedale River. Numbers on river represent gradient index(GI/I).12 cross-sections have been sketched from stereoscopic airphoto study (A to L).

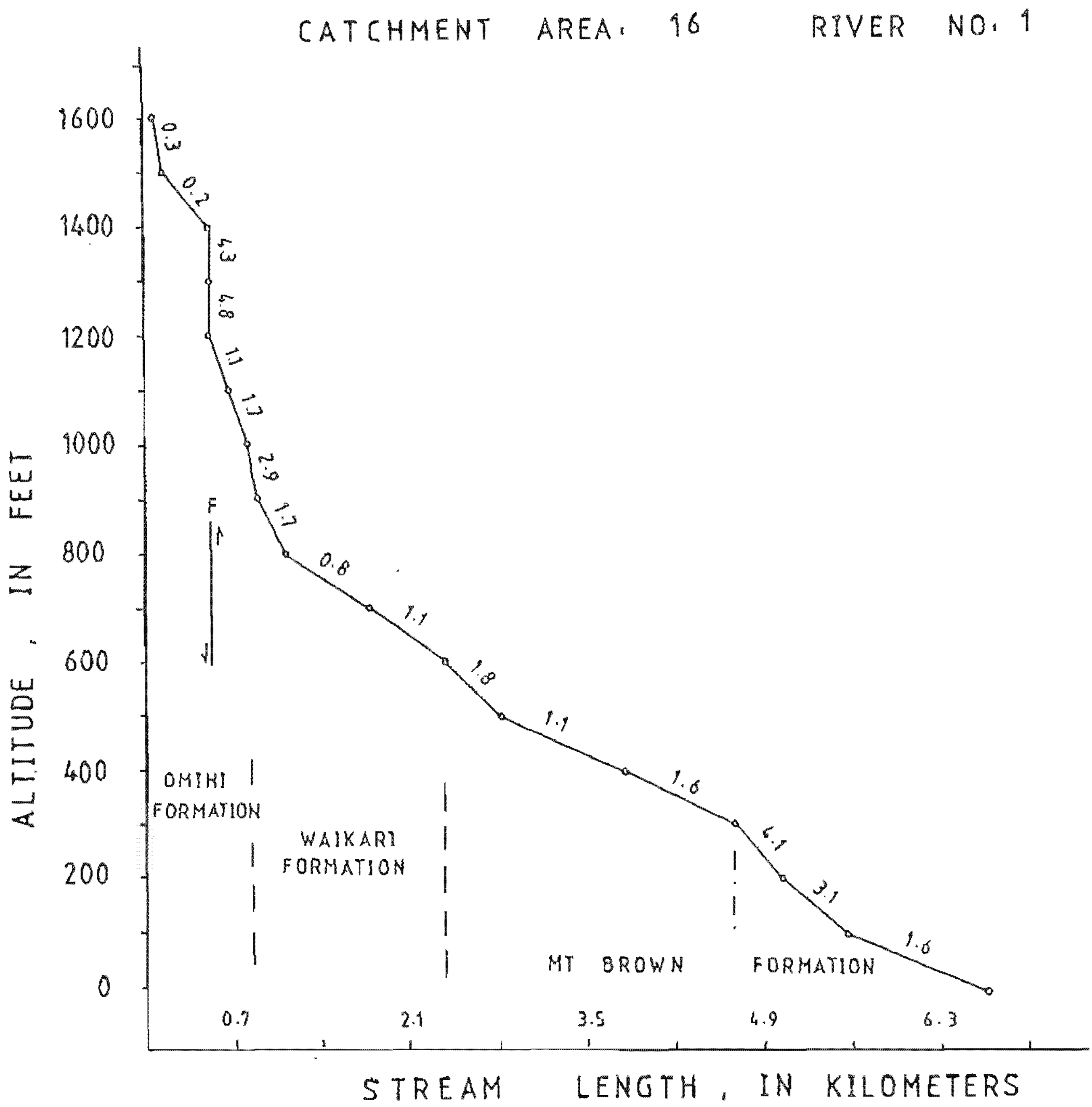


Figure 5.7 Longitudinal profile of the Dovedale River, Catchment 16. Numbers on river represent gradient index (GI/K).

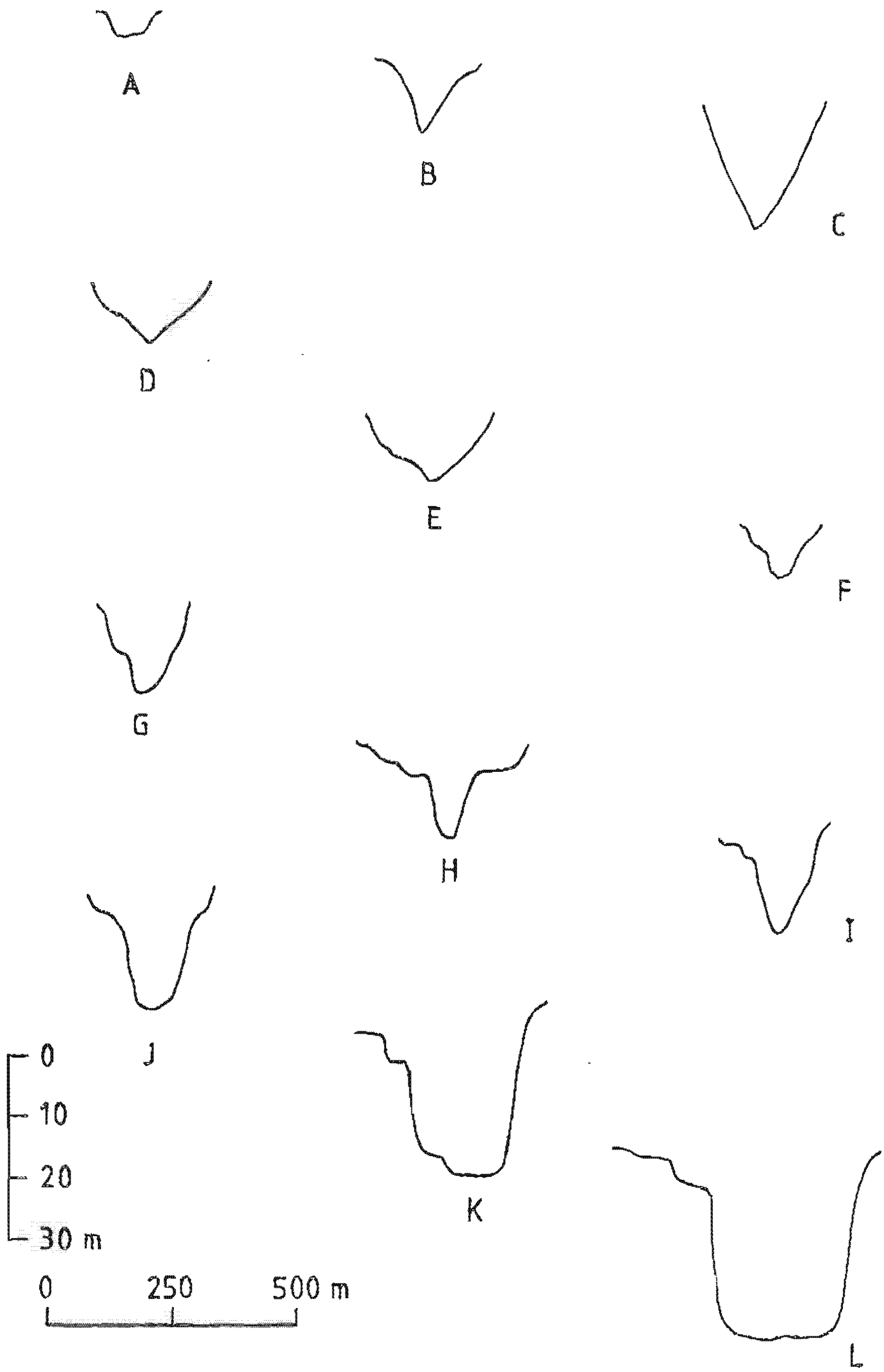


Figure 5.8

Cross sections of Dovedale River sketched from stereoscopic airphoto study lithological variation coincides with the major knickpoint identified in the Dovedale River at 300 ft. In this case the knickpoint has developed from a eustatic lowering of the base level in this case accompanied by differential uplift.

Regional stream rejuvenation has initiated a cycle of rapid downcutting. Rejuvenation by uplift is evident in the deep incision of the present drainage below the marine terraces (Plate 2).

The knickpoints of origins (a) and (c) differ from (b) in that they migrate upstream with time whereas, the lithologic change in gradient is controlled by the intersection of the incising stream with lithologic boundaries. Thus, types (a) and (c) will depend on their position for two controlling factors, time since initiation of disturbance and migration rate of knickpoint.

The profiles of channels that cross the moderately inclined structurally controlled hogback in the Weka Pass Stone, exhibit anomalously high gradients at the fault trace. This is well illustrated in Fig. 5.6. A new, previously unmapped, fault line can be recognized easily using the gradient index analysis. Thus the most consistent feature of these channels is the correlation between the structurally controlled hogback and the base of the zone of increased gradient. Knickpoints often coincide with the trace of faults. On the other hand, many inconsistencies appear when the correlation between river gradients and lithology is examined in detail, but some of the discordance is attributed to internal slumping of the limestone and many sink holes were observed (Plate 2).

Gradient index (GI/K) plots of the longitudinal profiles of the Dovedale's tributaries show several clear contrasts between the Dovedale's north and south bank tributaries (Fig. 5.6). The greater

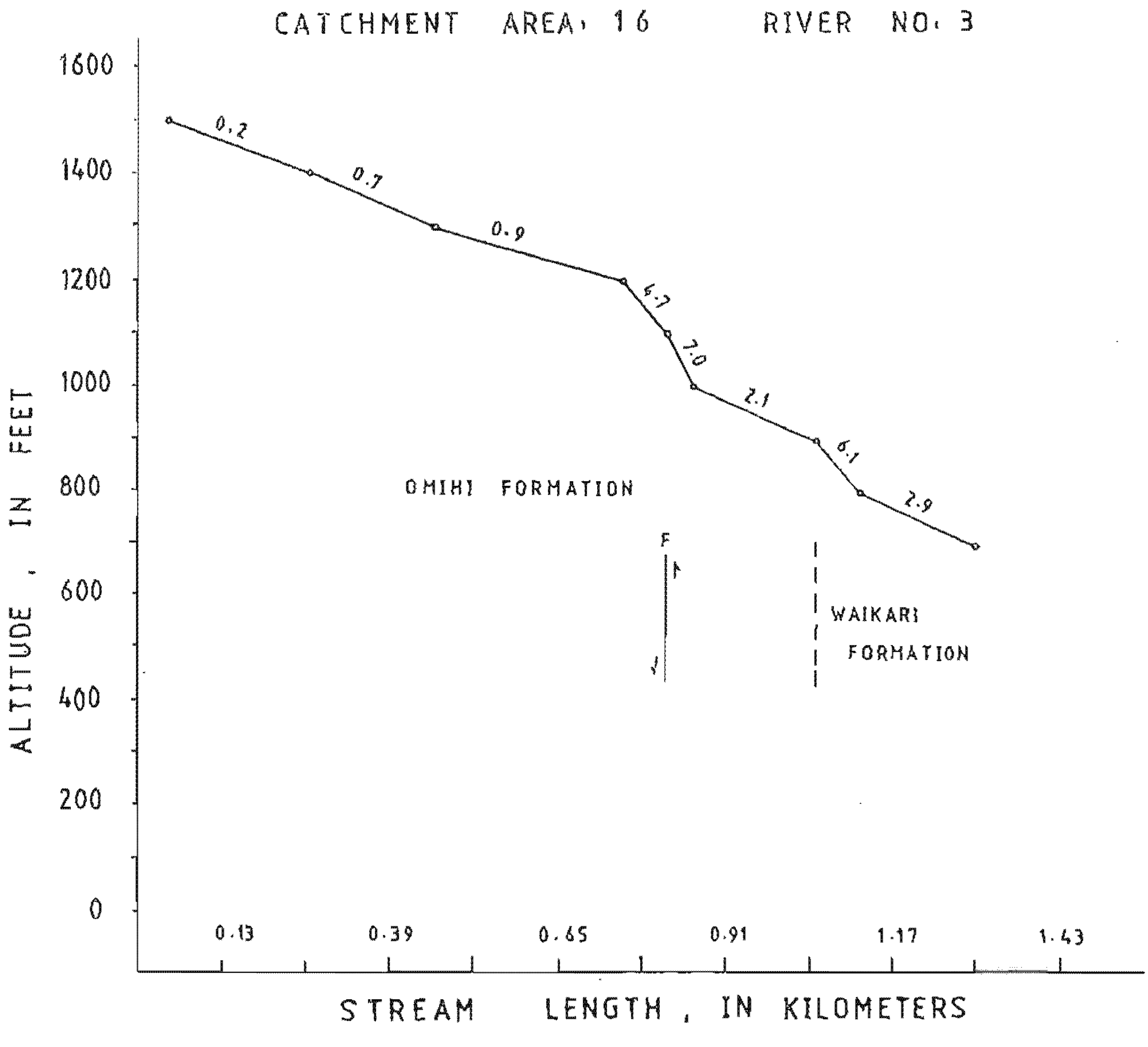


Figure 5.9 Longitudinal profile of the Dovedale's north tributary (river no 3).

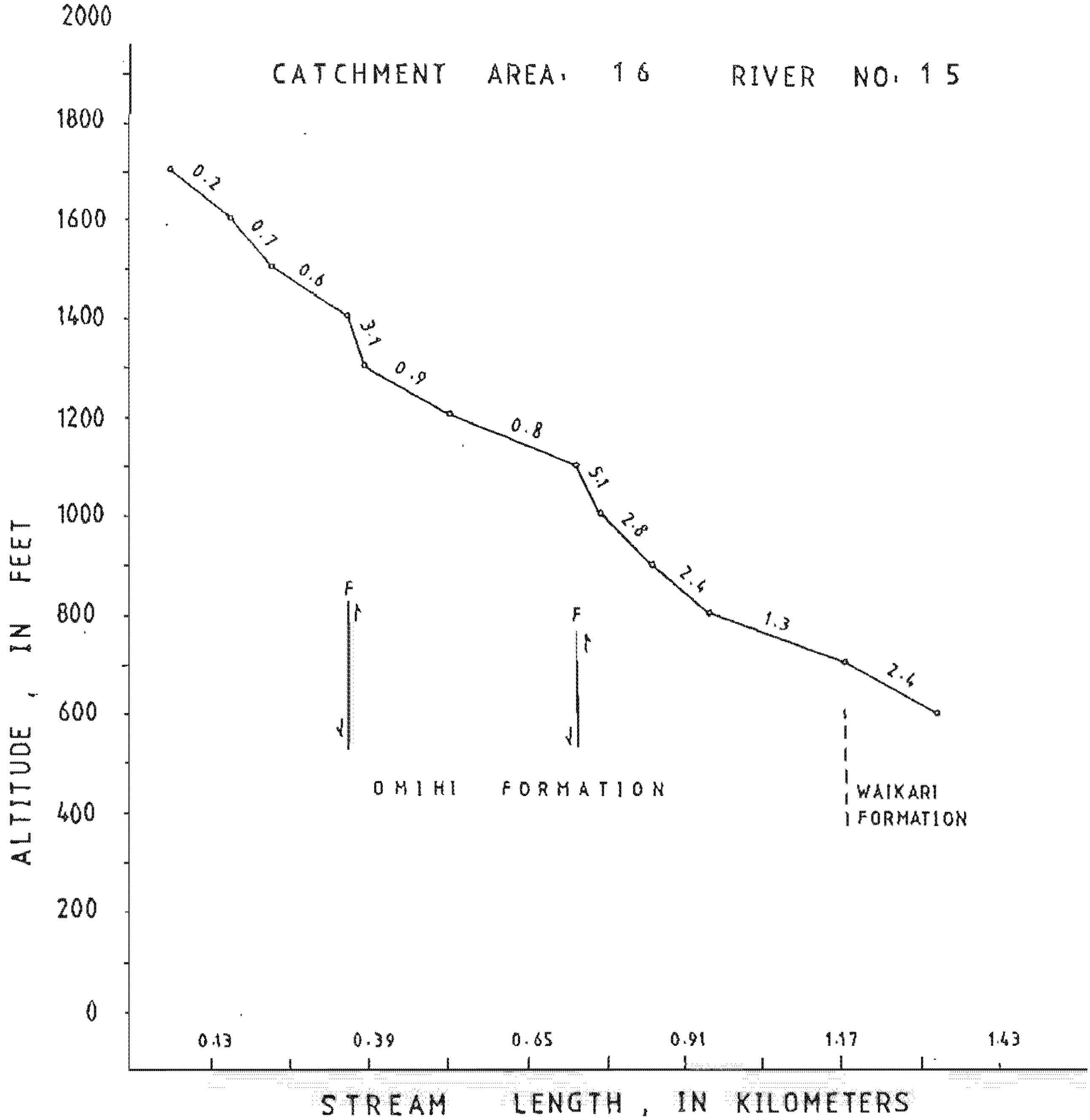
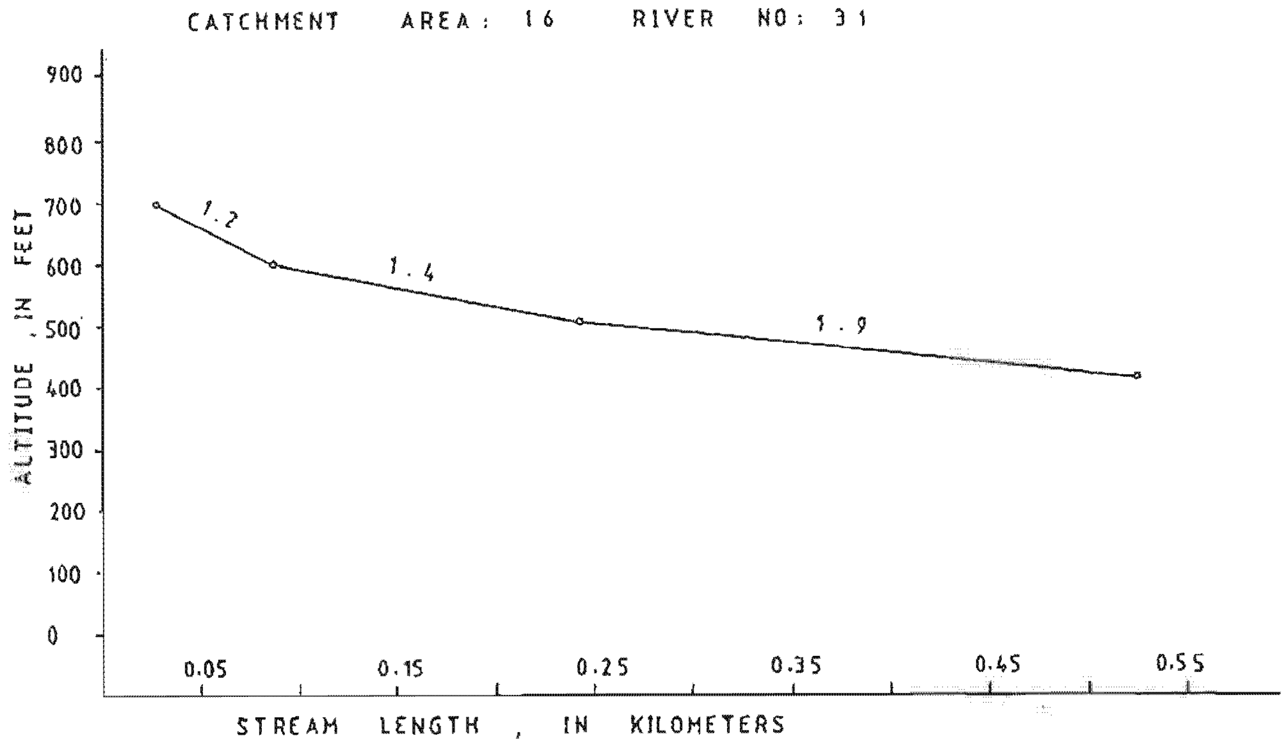


Figure 5.10 Longitudinal profile of the Dovedale's north tributary (river no. 15).



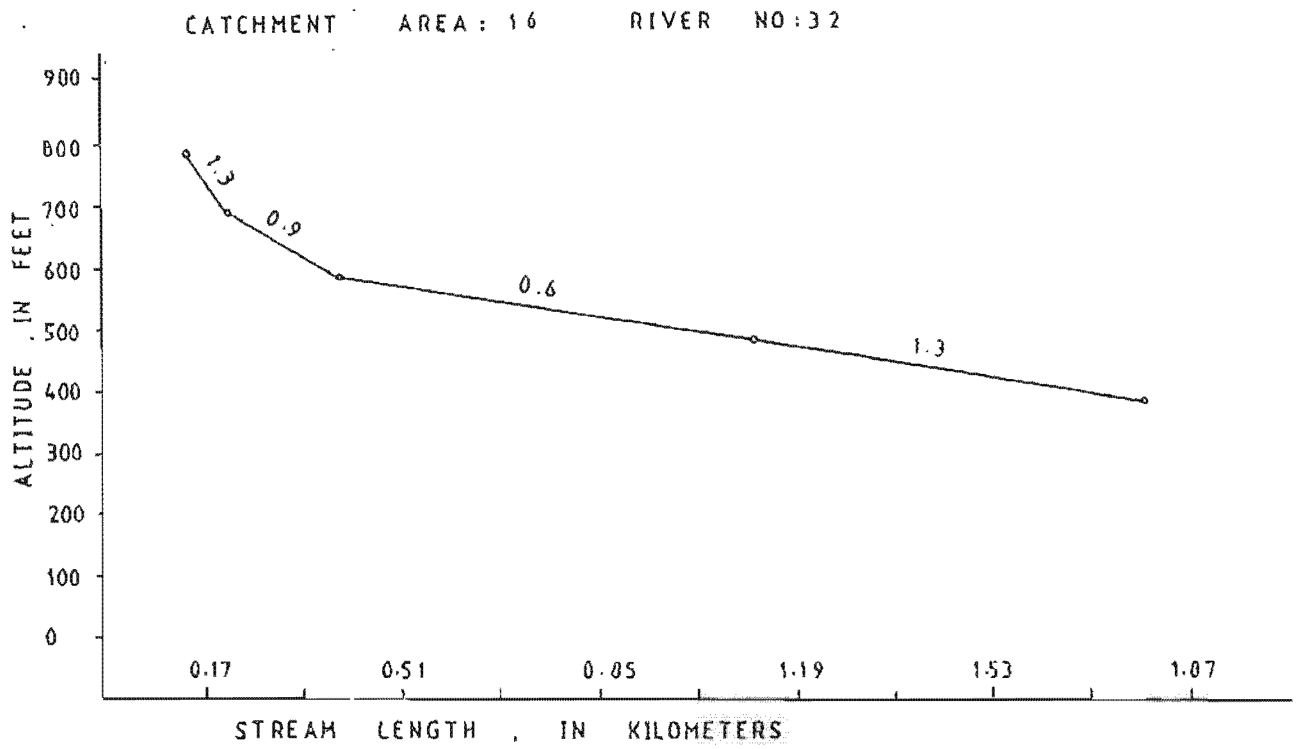


Figure 5.11 Longitudinal profiles of the Dovedale's tributary (river nos. 31 & 32).

number of major north bank tributaries is a reflection of the greater relief and steeper slopes which are generally found north of the river and which contribute to greater runoff. The south bank tributaries (Fig. 5.11) exhibit smooth, straight to gently concave profiles which make shallow junctions with the Dovedale River. The north bank tributaries (Figs 5.9 & 5.10), are steeper and more irregular throughout their profiles, especially immediately upstream of each fault. In this case, the high gradient index (GI/K) zones are associated with the en echelon faults along the structurally controlled hogback (Plate 2).

Three conclusions can be drawn from the profile of the Dovedale River: (1) that part of the profile above the 300 ft knickpoint is adjusted to the geology of the basin; (2) only the lower course has apparently been affected by changes in base level and uplift during the latest Pleistocene and Holocene sea level changes. This is an important point to establish when distinguishing eustatic from tectonic effects on base level; (3) the gradient index can provide insights into the causes of the diversity of the landscape.

5.5.2 DESCRIPTION OF CATCHMENT 27

The studies were made on about 29 streams in the area shown on Fig. 5.12 encompassing the catchment of the Teviotdale River. The structure of the area consists of an anticline and syncline. In general this anticline is directly related to the topography as a topographic high which descends from 1190 ft in the crest to about 200 ft at the trough. The syncline is also directly related to the topography as a topographic low (Fig. 5.13). Drainage patterns developed in this area indicate asymmetry. In general, the drainage

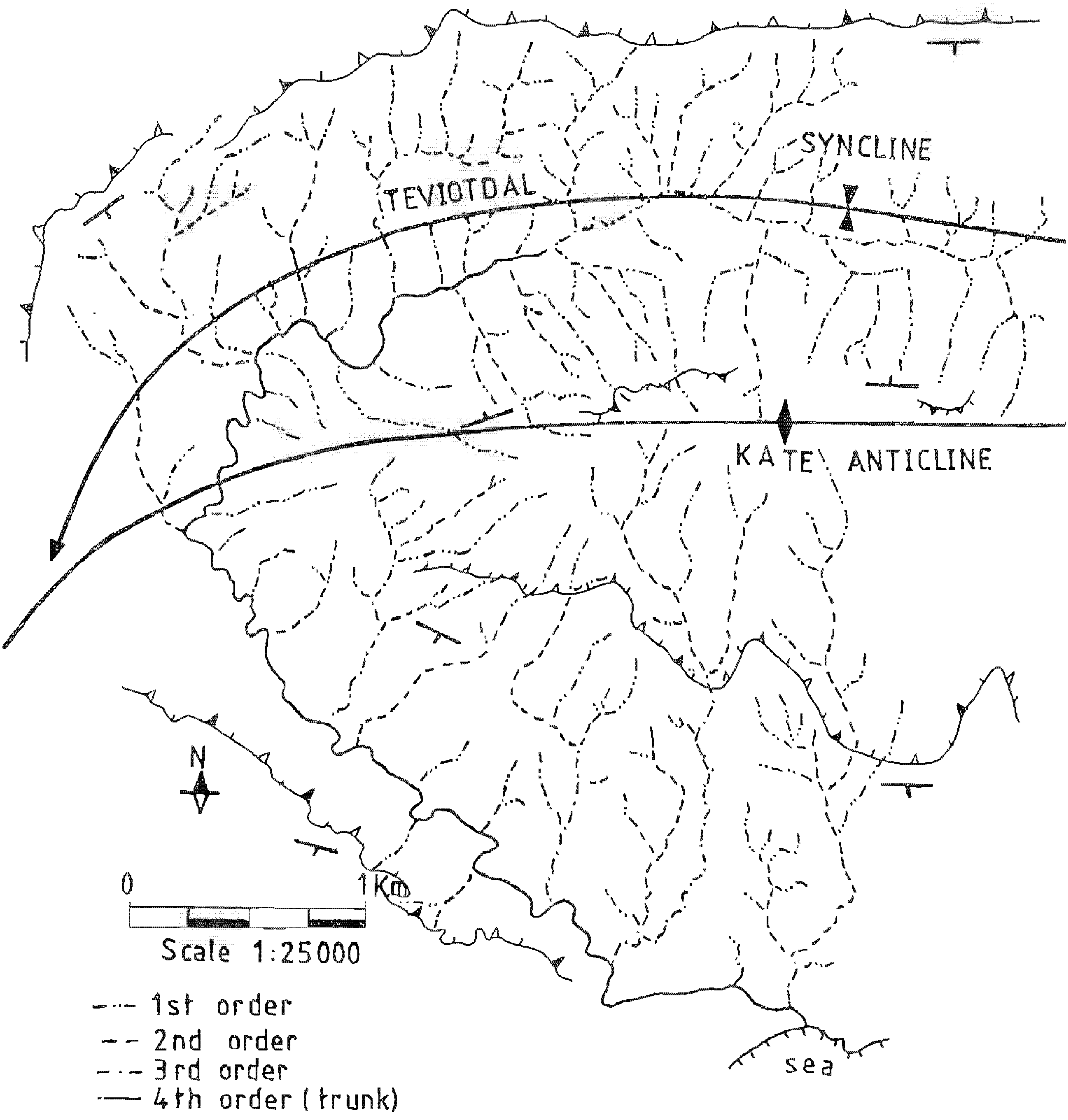


Figure 5.12

Map of the Teviotdale River basin in Waipara Region.

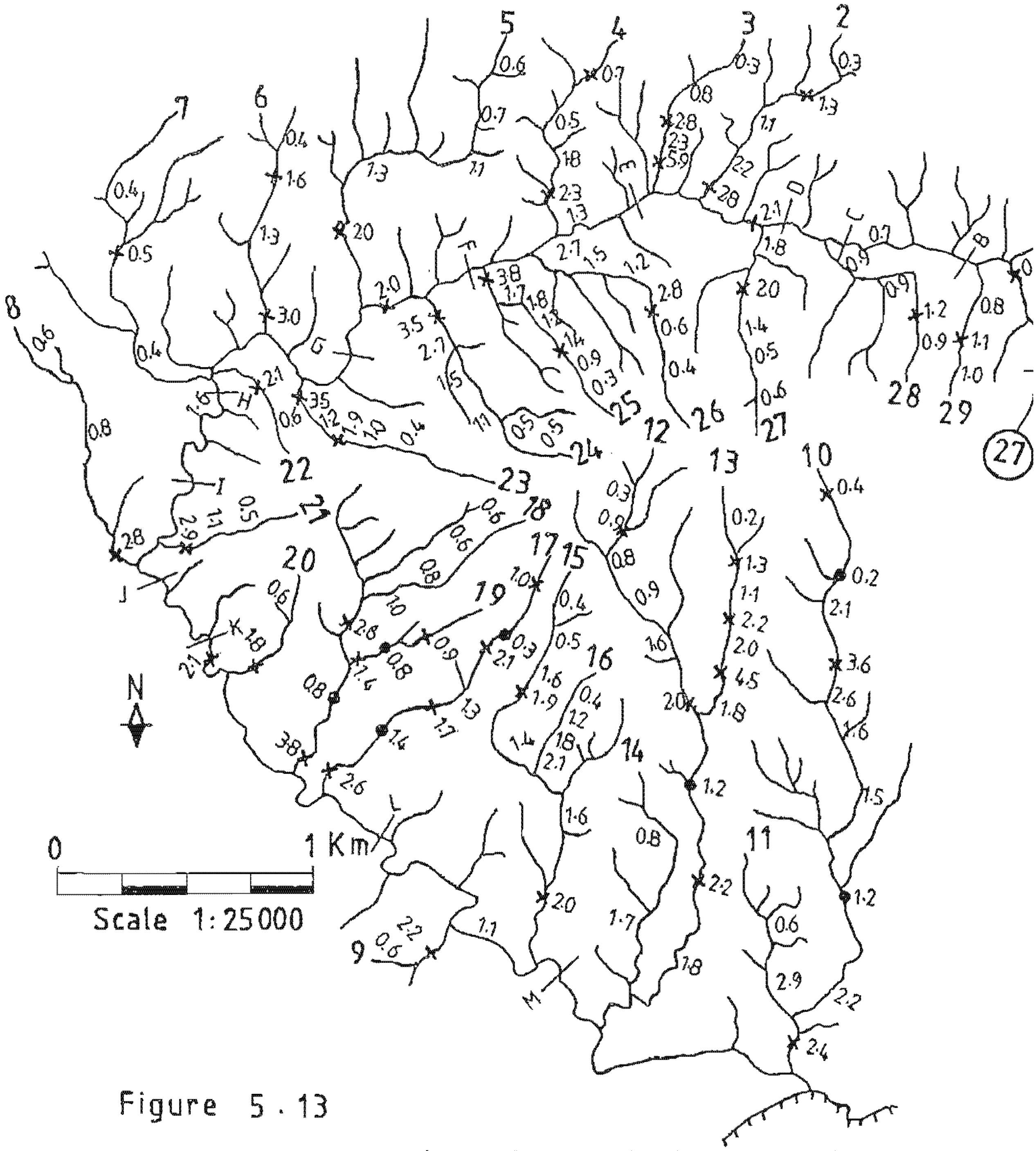


Figure 5.13

Drainage basin for Teviotdale River. Numbers on river represer gradient index(GI/K). 13 cross-sections have been sketched from stereoscopic airphoto study(A to M).

is parallel dendritic and indicates pronounced slope and structural control. The many tributaries joining the Teviotdale River are of resequent and obsequent nature while the Teviotdale River is probably hybrid in character as discussed below. The drainage is generally up to third order but in places exhibits only second order branching indicating an early stage of development; conversely the drainage texture is fine to medium, indicating maturity. These sorts of contradictory characteristics indicate inequilibrium conditions due to continuous tectonic movements.

The Teviotdale River drains about 7.7 kms of country underlain largely by Greenwood Formation and Tokama Siltstone. As shown in Figure 5.13, the longitudinal profile is composed of two distinct sections. In its upper reaches this river flows through a strike valley that developed along a synclinal axis (Teviotdale Syncline). In its lower reaches this river flows across an anticlinal axis (Kate Anticline) and then through a deep gorge. The Teviotdale River exhibits knickpoints around an altitude of 600 ft, 400 ft and 200 ft. (Fig. 5.14).

The first knickpoint at 600 ft, can be ascribed to lithological change, although this change is within the same formation (Greenwood Formation). There is a diversity of lithotypes, and this knickpoint occurs where there are massive resistant conglomerate beds. Here there is an abrupt change of cross-section from smooth cradle shape to sharp V-shape well-illustrated in Fig. 5.15 (B,C and D). Active downcutting by the stream is clear evidence that the threshold is now being exceeded. But stream power decreases with increasing distance from the headwaters and this is shown by the presence of alluvium in amounts in excess of that scoured by large discharges.

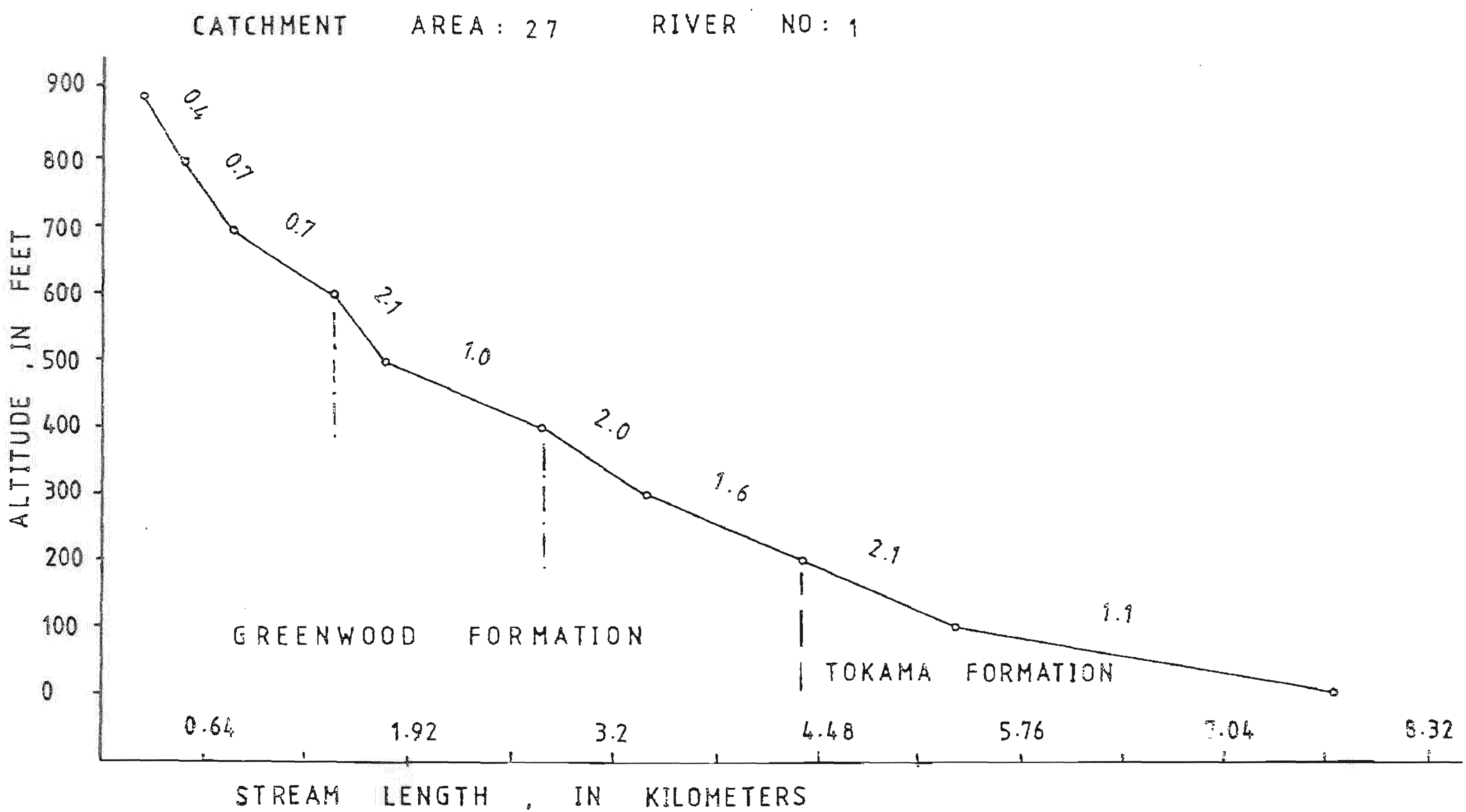


Figure 5.14 Longitudinal profile of Teviotdale River, Catchment 27. Numbers on river represent gradient index (GI/K).

Inconsistencies in the expected relationship between stream power and steep river gradients, accompanied by abrupt changes of the river cross-section described above, suggest that the role of differential uplift and lowering of the base level is secondary to the role of the lithological change in shaping the profile of Teviotdale River.

The second knickpoint 400 ft, probably can be ascribed to differential uplift. The behaviour of the Teviotdale River at the point where it crosses the axis of the Kate Anticline is an example of a tectonically induced anomaly. Two lines of evidence supported this idea. (1) The local peculiarity of the meandering course, is an interruption of the normal pattern, probably caused by the appearance of the growing anticline along its path, an important phenomenon discussed more fully in Chapter Six. (2) Distinct changes in cross-section of this river on either side of the anticline suggests structural control. Local narrowing of the valley and deep incision (see Fig. 5.15, H and I) produced by a differential uplift influenced the rate of valley narrowing. A shallow cradle cross-section was recognized (see Fig. 5.15, J) downstream of this incision. The significance of this is thus a suggested antecedent origin of the gorge which will be discussed in detail in the next chapter.

The third knickpoint 200 ft, can be ascribed both to lithological change and to differential uplift. This knickpoint was observed in the area where there is a formation boundary i.e. between Greenwood Formation and Tokama Siltstone.

Steep sections on the gradient index plots which cannot be explained by lithological control may be taken to indicate a disequilibrium in the profile, most certainly due to relative lowering of base level (Hack 1973b). This steepening should also be present in

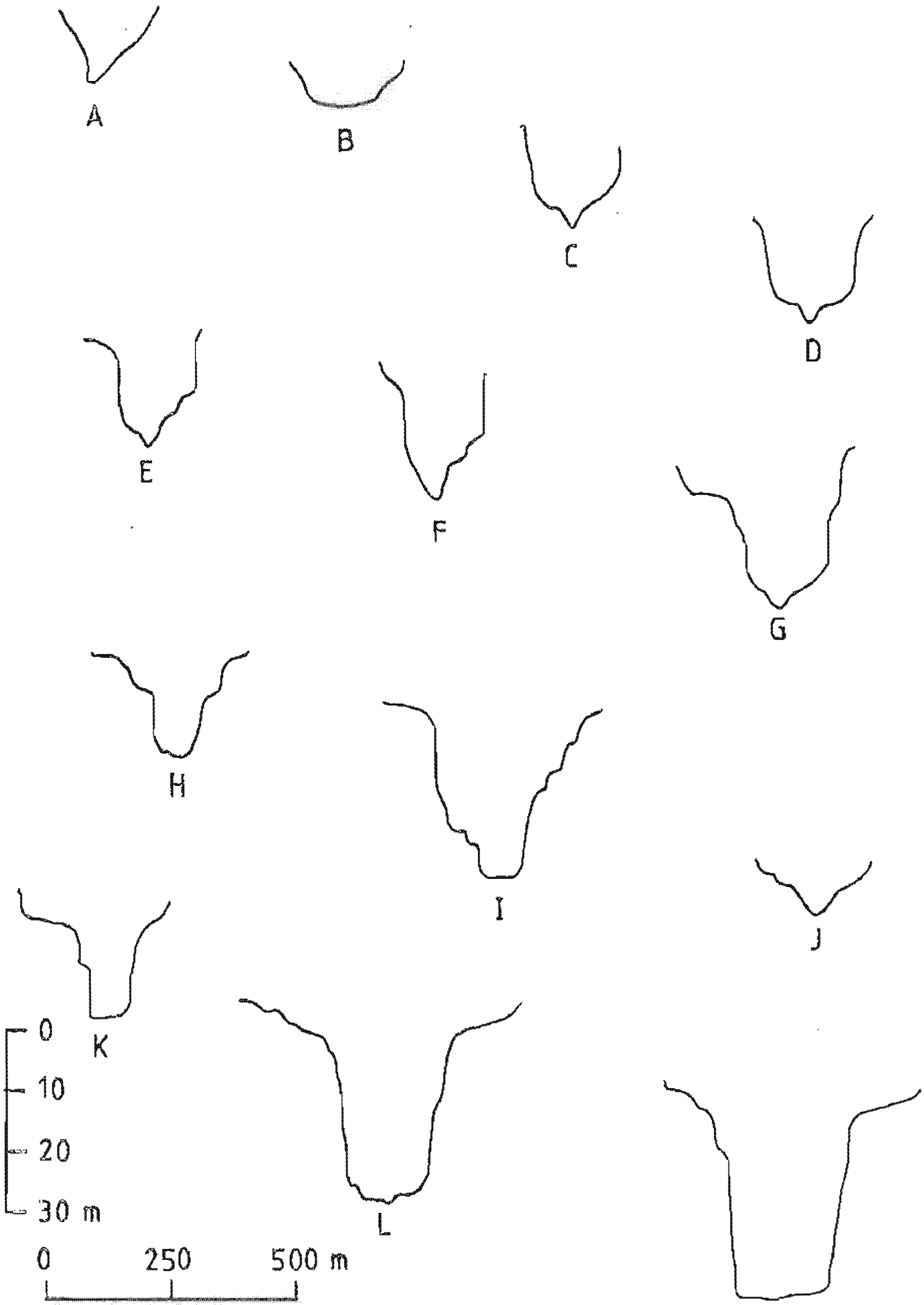
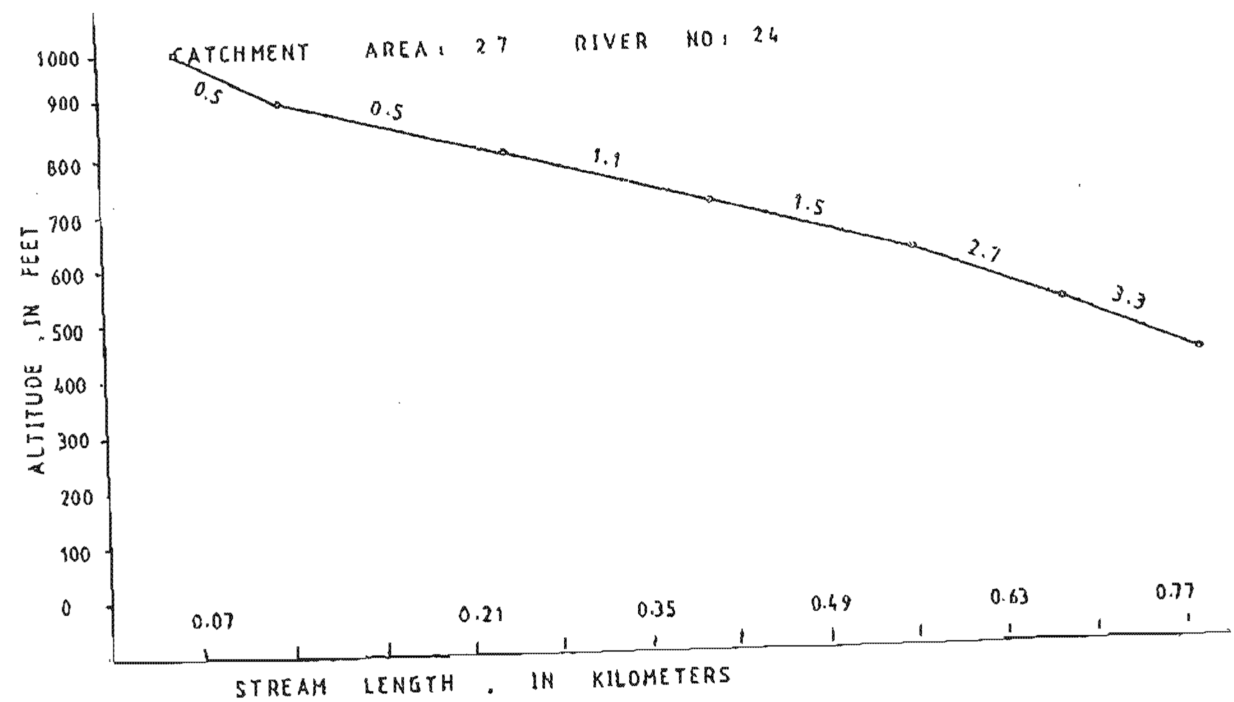


Figure 5.15
Cross section of Teviotdale River sketched from stereoscopic airphoto study.

tributaries which enter a trunk stream within or below any oversteepened reach which is not explicable by lithological control. An example of such oversteepening is provided by the Teviotdale River upstream from the meandering valley. Figure 5.16 has shown that the north and south bank tributaries below the 600ft knickpoint enter the Teviotdale River with steep gradients on the gradient index plots (upper profile). Above the 600ft knickpoint both north and south bank tributaries have smooth junctions with the Teviotdale River (lower profile). Thus, the deep gorges at the mouth of catchment 16 and 27 are also attributed to the relative lowering of base level (i.e. sea level).

For the central and southeastern margin of this catchment, there is a fairly direct relation between the differential uplift and the main physiographic features. Streams flowing across the area are now entrenched. The Teviotdale River had downcut over 100 ft (30 m) into Tokama Siltstone. As previously mentioned, rejuvenation by uplift is evidenced by the steep incision of the present drainage below the marine terraces and underlying formations.

Three conclusions can be drawn from the profile of the Teviotdale River. (1) The presence of a distinct knickpoint along the 400 ft suggests that the Teviotdale River is maintaining a course antecedant to a rising anticline and is keeping pace with differential uplift. (2) The lower course has apparently been affected by changes in base level and regional uplift during the Pleistocene and Holocene time. (3) A change of channel cross-section and meander characteristics can provide a clue to indicate areas of active uplift and to changing base level. Thus, the longitudinal profile of a river is sensitive to ongoing processes of the uplift and can be used to



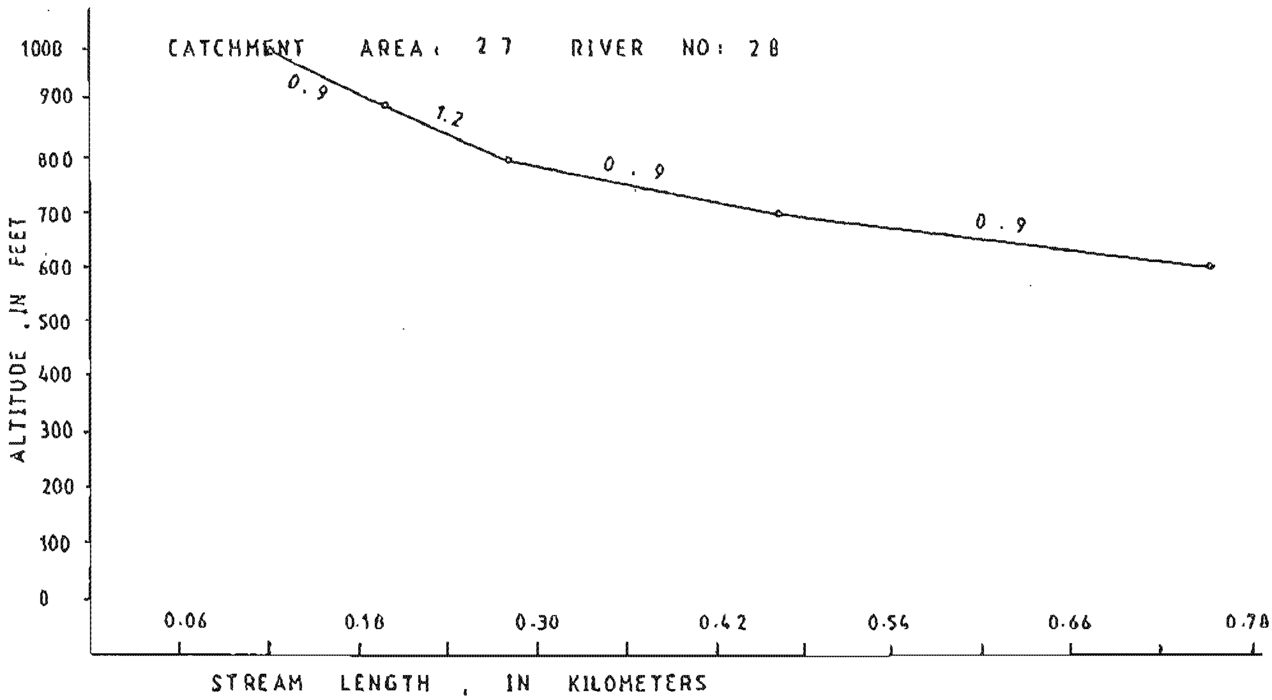


Figure 5.16 Longitudinal profiles of the Teviotdale's tributaries (river nos. 24 & 28).

recognize active structures.

5.6 MARINE TERRACES AS A RESULT OF EUSTATIC SEA-LEVEL CHANGE AND TECTONIC UPLIFT

An extensive flight of uplifted marine and fluvial terraces on the eastern slopes of the coastal range of the Waipara region provides a series of datum levels for determining the rate of a number of geologic processes. The relative ages of sea level stands can be used to develop models of late Quaternary deformation and geomorphic evolution.

Often there is a need for age determination of late Quaternary marine terraces, but little or no material is available for radiometric dates. Despite the humid climate, samples for radiocarbon dating are not common in both Teviotdale and Canterbury aggradation gravels and the higher marine surfaces may be expected to be older than the limits of this dating method. Therefore, the former shoreline height is used as the best criterion for the relative dating but requires considerations of the vertical deformation due to tectonic or isostatic origins in the coastal area during the late Pleistocene before correlation with dated sequences.

Volkov et al. 1967 were perhaps the first to quantify effects of recent crustal movements on the shape of longitudinal profiles and water levels in rivers. To evaluate the relative contribution of tectonism to a break in the longitudinal profile, a theoretical profile is calculated. Unfortunately, the reference does not deal with how the theoretical profiles accomplish the filtering of tectonic activity. Apparently a profile is generalized from the existing profile data. The conclusion of this study was that maps could be compiled characterizing the summary amplitudes of neotectonic movements in the Holocene.

Mayer (1979) assumes that for uniformly dissected terrains, each rock type will have its own characteristic slope. In order to separate the component of channel steepening contributed by the resistant strata, and the component of steepening from tectonic alteration of the channel, Mayer uses a modified gradient index formula of Hack (1973a). This method has not been given a thorough test. Mayer (1984) recently discussed the complicating effects of measurement error, lithology, particle size, and composite scarps.

Hanks et al. (1984) examined a single shore-normal profile of marine terrace sea cliffs cut into the Pliocene Santa Cruz Mudstone. They used terrace ages based on a uniform rate of uplift in order to derive estimates of the "mass diffusivity" (which is part of the diffusion equation model of slope evolution) for each terrace; their estimates of this parameter for two high terraces agreed with their calculated value for a better dated low terrace. The conclusion of this study was that ages derived from the constant uplift rate assumption were reasonable.

Crittenden & Muhs (1986) proposed another method using the log-linear relationship of Bucknam & Anderson (1979), between scarp height and slope angle using sea cliff height and maximum slope angle. These investigators examined a chronosequence of marine terraces cut into Miocene andesite on San Clemente Island, California. Overall, the relationships are weaker than for stream terraces and fault scarps in unconsolidated materials, but the method can probably be used successfully to distinguish early, middle and late Quaternary fault scarps or sea cliffs in consolidated materials.

However, no study that I am aware of has tested the stream gradient index formula of Hack (1973a) on undated marine terraces of late Quaternary age. Such a test is important, because gradient indices tend to be fairly constant for a given lithologic and climatic setting. Gradient index values indicate the steepness of a channel, "normalized" with respect to discharge. Uplift tends to increase slopes and a change from glacial to interglacial conditions has also increased stream slope in response to eustatic sea level change. The nature of the channel steepening must be investigated before attributing the high or low gradient index value to any one or more factors.

Gradient index (GI/K) values were calculated for reaches along most of the coastal channels from headwaters to the sea (Appendix 1). Gradient index for particular reaches higher or lower than the mean value are displayed in Plate 4. Knickpoints on streams are controlled by lithologic boundaries, resistant members, fault lines, postglacial cliffing and incision and uplift of former shorelines. Having geologic and geomorphic maps enables one to examine GI/K values that are high or low and which cannot be explained by rock type. The dilemma is, how much of the GI/K value is due to the tectonic warping and uplift steepening the channel, and how much is eustatically induced. There does not seem to be a simple method of separating these two components and for this area it is evident that both are contributing factors.

On a slowly tectonically rising coast which is experiencing a high stand of the sea, a sea cliff is undercut as the shore platform extends inland, and a new marine terrace is formed when the sea retreats below the platform at the end of that interglacial stage. After deformation (in the case of growing folds) or sea level returns to low stand, stream profile slopes are subjected to knickpoint migration processes that alter the initial slope through time. Although not all points along a given wave cut erode at once, little error is introduced by regarding all points along wave cut to be essentially synchronous because the rates of tectonic uplift are slow relative to the time occupied by pauses in eustatic sea level change. Thus the series of knicks work headward with progressive entrenchment below. The spacing between knickpoints depends on the magnitude of the uplift, the rate of incision into various rock types and discharge differences along the stream and the time between each successive high stand.

Assume that while continuous folding is taking place, a horizontal surface is cut instantaneously at sea-level. The lowering of the base-level creates a knickpoint and initiates an incision which migrates upstream maintaining a sharp knickpoint. If successive surfaces are cut at sea-level, the contours of each surface will show the same fold axial traces, but the absolute contour values and gradients of each surface will depend on the age of the surface. Consequently the several surfaces can be interpreted as rates of vertical movement by dividing the height of each by the age of each bearing in mind the potential for a eustatic correction factor for any sea level stand that might be correlated with a lower sea level position than the present (Bull 1985). It follows that the distribution of the knickpoints also define surfaces which are related to a wave-cut surface and the present heights of these knickpoints are due to the combined effects of eustatic height and tectonic uplift plus a component of headward migration.

The elevation and the exact position of these knickpoints on the ground is determined either by photogrammetric techniques or various standard surveying methods to reduce the error in calculating the elevations from the topographical map (Plate 3). Potentially comparison of these elevations with those of the corresponding marine terrace would allow the relative rate of migration of the knickpoints to be determined. In practice the knickpoint migration rates appear to be slow and the fact that along this section of coast the streams are of similar size and cut in similar lithologies has meant that any differences in elevation between knickpoints is negligible.

5.7 HOW OLD ARE THE TERRACES?

Plate 2 shows well developed tilted marine terraces which are cut into soft Late Tertiary sedimentary rocks on the eastern slopes of the coastal range of the Waipara region. They are M, M1, M2, M3, M4 terraces respectively from youngest to oldest. There is no evidence for determining the ages of marine terraces on the study area, other than the local preservation of Kawakawa ash on the coast near Teviotdale, which is 20,300 + 7,100 yr. B.P. (Kohn 1979) and the inferred age of the Canterbury surface 10,550 + 150 yr. B.P. (Harris 1982). Nevertheless their relative ages can be unambiguously inferred from their relations to multiple periods of aggradation (Teviotdale and Canterbury Surfaces) and estimates of their approximate ages deduced by correlation.

The Canterbury surface is truncated at the mouth of the Waipara River by MO but is itself incised into M1. Ancestral Waipara River terraces are clearly confined to the ancestral Waipara valley incised into Teviotdale Surface and M1. These relationships are clearly expressed on the map (Plate 2). The Teviotdale Surface is truncated at the coast by Mo and M1 but is itself incised into M2. The age of the older marine terraces can only be estimated indirectly. The raised marine terrace sequence throughout the study area is morphologically similar to a well developed sequence described from the Conway Coast by Ota et al. (1984). These investigators proposed that the highest marine terrace would have formed 125,000 years ago during the last interglacial sea level maximum, an estimate compatible with the average rate of uplift in late Holocene time.

Present altitudes of marine terraces are determined by local rates of uplift and by the altitudes of global glacio-eustatic high stands of late Pleistocene sea level. The chronological framework for this late Pleistocene period is based on climatic fluctuations from glacial to interglacial conditions. Duplessey (1978) presents a late Quaternary time scale (Fig. 5.17) which is based on oxygen isotope records. Dates for the New Zealand last interglacial boundaries are from Suggate (1974) and isotopes stages 1 to 7 are from Emiliani (1972), and dating of eustatic high sea level stands by correlation with the dated New Zealand and Papua New Guinea sequence (Chappell 1974, 1983 etc).

Marine terraces of supposed last Interglacial age (oxygen isotope stage 5, c. 80,000- 130,000 years B.P.) are recorded from many locations in New Zealand (Chappell 1974, Ota et al. 1981, Pillans 1986, Ota 1987). As the 125 ky high sea level has been assigned eustatic elevations corrected for isostatic and tectonic components from both stable and uplifted areas, the accepted range from +2 to +10m is now well established. A value of +6 is generally accepted for use in tectonic studies (Stearns 1976, Chappell 1975). Two other high

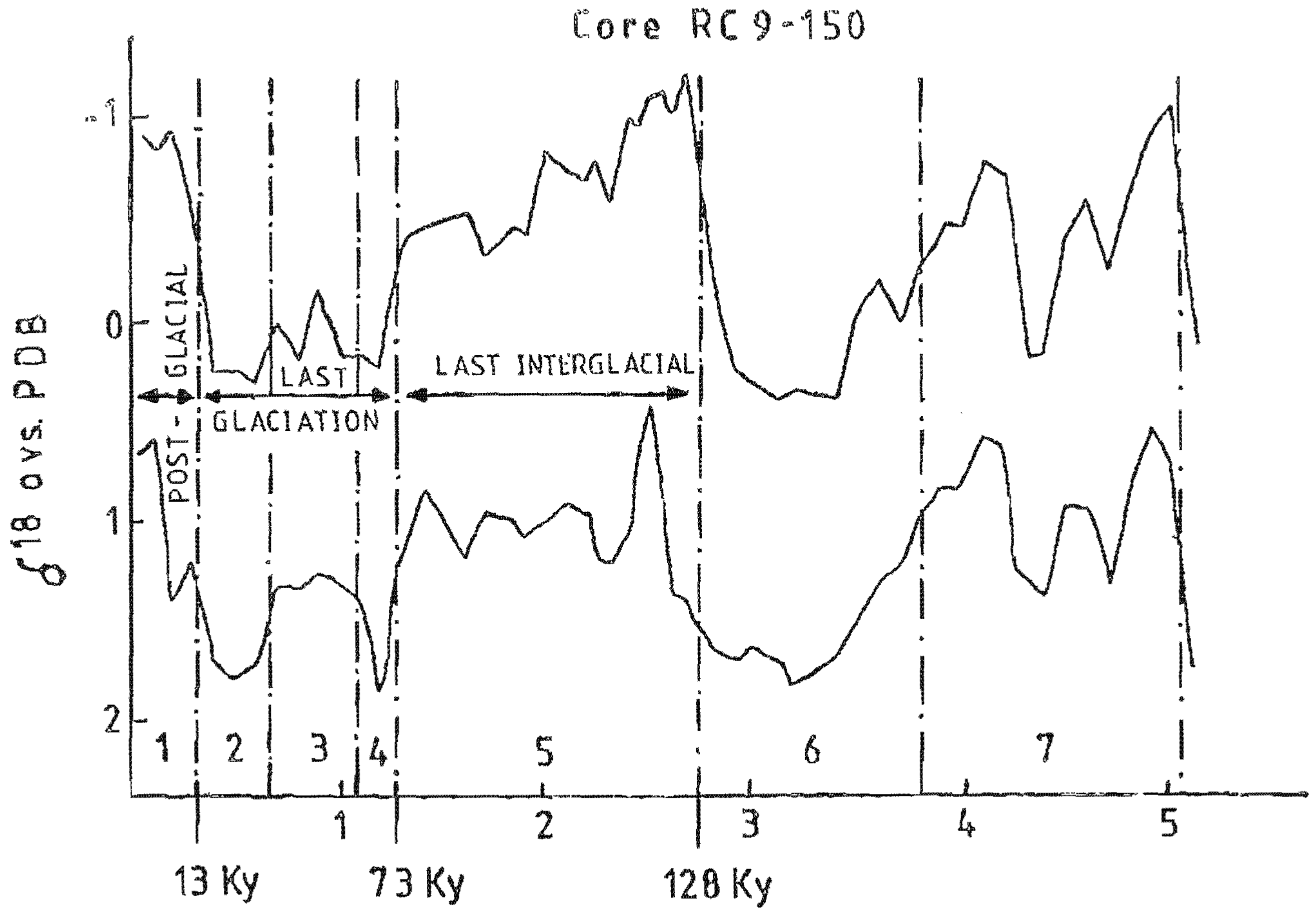


Figure 5.17

Oxygen isotope record of the planktonic species (a) Globigerinoides sacculifer and G. inflata in the Indian core RC 9-150 (31°17 S, 114°36 E, 2703m). Isotope stages 1 to 7 are from Emiliani (1972). Date for the New Zealand Last Interglacial boundaries are from Suggate (1974). (Figure 1 from Duplessey (1978) fig. 3.4) sea levels of the last interglacial at c. 80,000 and c. 105,000 years B.P. are also well documented overseas, when sea level lay between c. -10 and -20 relative to present (Bloom et al. 1974). A value of -13 and -15 m is generally accepted for use as a correction factor in tectonic studies for the two stands respectively. The three major high sea level events described above at c. 80,000, 105,000 and 125,000 years B.P. are correlated to oxygen isotope stages 5a, 5c and 5e respectively (Shackleton & Matthews 1977) separated by minor temperature drops (Fig. 5.17).

The last glaciation is probably bracketed by the ages of c. 73 ky and c. 13 ky (Shackleton & Opdyke 1973). Two warm interstadials dated at c. 60 ky and c. 40 ky (Chappell 1974) separate three cold stadials.

A rhyolitic ash was preserved within a thick loess unit, which occurs at 47 m a.s.l. (i.e. 7.6 m below the Tiromoana coastal plain M1), contains fresh glass shards which gave a fission-track age of 20,300 +/-7,100 yr.B.P. (Kohn 1979). The sediment containing the ash was inferred by Carr (1970) to be Oturian to early Otiran in age suggesting an age range between c. 50,000 and 130,000 yr.B.P. (Fleming 1975). This apparent contradiction in the evidence for the age of the Tiromoana coastal plain is reconciled by Kohn's observation that the ash actually lies within the uppermost of the two loess horizons, separated by a weak paleosol, which blankets the original marine surface. Harris (1982) correlated the episode of aggradation forming the Canterbury Surface in the Waipara-Omihi valley, with deposition during and since the Poulter advance of the Otira Glaciation dated by 14C, obtained from a Hyridella sample collected from a shell bed in the uppermost silt unit, at c. 10,550 +/-150 yr.B.P. But variability of Hyridella 14C standards with locality and time make it unreliable for dating (Currie 1984). A radiocarbon date from the Cass district shows that the Poulter event ended some time before 13,750 +/-200yr.B.P., but the time is not closely pinpointed (Burrows 1983). Nevertheless, this local date must be regarded as a provisional minimal estimate for the end of the Otira Glaciation. The Canterbury Surface along Waipara River is a little lower than the Tiromoana coastal plain (M1) at the river mouth and can be traced upstream to the Waipara-Omihi valley. The tephra is therefore coeval with the penultimate advance of the Otira Glaciation, but has not been identified on the Canterbury Surface, thus supporting the estimation that locally the Canterbury Surface is related to the latest Otiran advance. While a 10,000 yr.B.P. minimum age for the Canterbury Surface has been adopted for estimating deformation rates. From this tentative correlation, the Tiromoana coastal plain is inferred to be the 60,000 B.P. marine terrace.

The heights of the complex of geomorphic knickpoints evident in the river profiles crossing the eastern coastal slopes can be divided into four separate zones (Plate 4). The correlation between knickpoints and the marine terraces is generally good as shown in Plates 2 and 4. This good correlation leads the author to utilize the profile data to extend the last Interglacial chronology throughout this study area. The four geomorphic zones correspond to the well known shorelines at 60, 80, 105 and 125 ky repsectively from youngest to oldest (Plate 4) and correlated ages of the key geomorphic datum surfaces discussed are summarised below.

Hence Terrace M0 is inferred to be Holocene

Canterbury Surface	10 ky
Ancestral Waipara Terraces	<20 ky
Terrace M1	60 ky
Teviotdale Surface	?70-75 ky
Terrace M2	80 ky
Terrace M3	105 ky
Terrace M4	125 ky

5.8 LATE PLEISTOCENE DEFORMATION DEDUCED FROM FORMER SHORELINE HEIGHTS

Plate 5 shows the former shorelines on late Pleistocene marine terraces projected onto a section parallel to the general trend of the present shoreline of the Waipara coast. The height and areal distribution of former shorelines vary markedly, indicating differential uplift. The former shoreline profiles are compared with geological and geomorphic maps, in order to obtain quantitative characteristics of the deformation. These can be considered as approximately equal to the incremented net amplitudes of vertical crustal movements during the late Pleistocene. It was found that the rate of uplift of the four levels of marine terraces is quite proportional to the relative change in rates around local structures crossed by the shorelines as shown on Plate 5. The appropriate eustatic correction is incorporated. The southwestward decrease of the heights of the former shorelines is compatible with the meridional trend of the major folds in the study area.

The rates calculated for the intervals of time between the formation of successive marine terraces, however, indicate that the uplift rate for the region as a whole has not remained constant through time. The pattern of deformation indicates the highest rates of uplift occur in the plunging culmination of Kate and Montserrat Anticlines area (2.08 mm/yr). This rate of uplift is comparable to other known rates of crustal uplift along the Conway Coast namely an uplift of 1.8 - 3 mm/yr for the Hawkeswood Range (Ota et al. 1984).

An extensive flight of marine terraces between Motunau River and Boundary Creek is offset by a reverse fault as shown on Plate 5.

In the area south of Dovedale River, the low rate is associated with the Teviotdale Syncline and the highest rates are found along those parts of the shorelines crossed by the sigmoidal fold axis of the Kate Anticline.

The relative height of sea level peaks and the interval betweeen them are of major importance. A newly formed terrace is likely to be destroyed by coastal erosion if it is narrow and has insufficient freeboard above a subsequent sea level peak, as happens at 105 ky years in the eastern limb of the Kate Anticline and at 60 ky years at the Motunau River mouth area. Had either this sea level peak or the uplift rate been higher, this terrace might have survived much better.

On the basis of a simple growing fold model, uplift rates would be expected to incrase linearly inland across the marine terraces which have been preserved along the limbs. On this basis, the shore-normal deformation pattern across the southern limb of Kate Anticline suggests that uplift appears to have been an intermittent process rather than uniform type of uplift pattern. The model is presented in (Fig. 5.18) in the form of a relation diagram, which is a plot of terrace height vs. distance for each projection of the shore-normal terrace levels. The progressive increase in cumulative

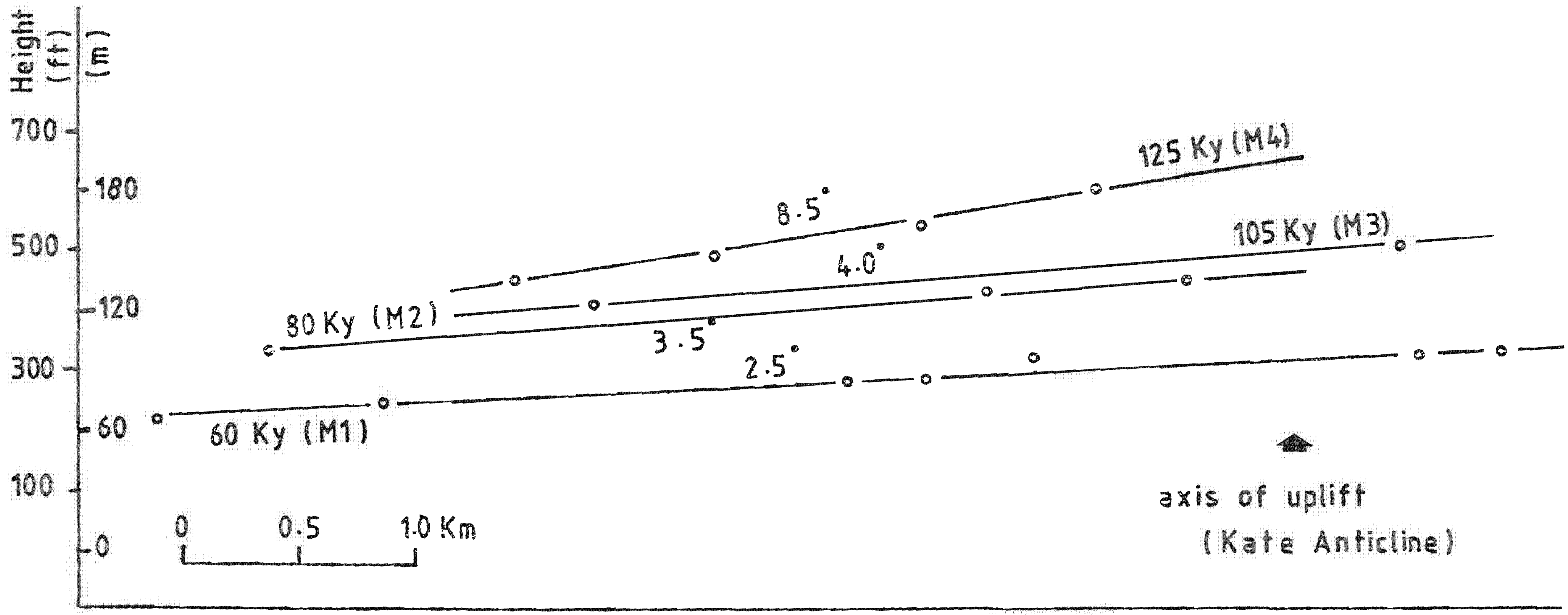


Figure 5.18 Shore-parallel projection of tilted marine terraces across the southern limb of the Kate Anticline.

tilt rates between the lower and upper marine terraces could be interpreted as a simple growing anticline. The gradation on the oldest marine terrace, 125 ky, is markedly steeper than the gradient that might be expected by extrapolation of the cumulative gradients observed on the 105, 80 & 60 Ky surfaces. If the correlation and ages suggested are correct, the uplift of the Kate Anticline has been most active during the period from the ages of 105 ky terrace to that of 80 ky terrace. This high uplift rate has a major control on the deviation of Kate Creek (see Chapter 6).

5.9 CORRELATION OF THE STREAM GRADIENT ANOMALIES TO GEOLOGIC AND GEOMORPHIC MAPS OF THE STUDY AREA: CONCLUSIONS

Along longitudinal river profiles we usually observe alternations of concave and convex river segments indicative of stream gradient anomalies. These deviations in the profiles depend on the direction of development of erosive processes. In the basin or break of slope area (for example Canterbury Surface) there is a tendency for river beds to fill with alluvium and a concave knickpoint sign together with a decrease in gradient index value were observed (see Appendix 1 catchment 33, 36, 37, 38, 39, 40, 49, 50, 51, 55, 56, 57, 58, 59). In uplifting areas scour of the river bed occurs and a convex knickpoint sign together with a marked increase in gradient index value correspond to the deformation zones. It is obvious that a change in the base level in the principal river involves a change in the intensity of erosive processes in the coastal range of the Waipara region. On dropping of sea level and vertical uplift in the coast, there occurs intensification of the erosive activity of small tributary streams and a more rapid development of gullies (see

Appendix 1 catchment 4, 5, 6, 7, 19, 20, 21, 62, 63).

The terrain has been cut from a variety of rocks under dominantly humid climates. The amount of incision by streams seems related to variations in stream power as well as rock resistance over both time and space. Many distinctive features can be seen in the comparison of the modern and old drainage rivers in the coastal region (e.g. Kate Creek [Catchment 18] and Motunau River [Catchment 3]). In detail , the two groups of profiles exhibit differences, the most important being the convexity of the modern valley profile profile compared with the general concavity of the old profile. See Appendix 1 for comparison between different catchment profiles. The convexity was discussed before and explained in terms of the effect on longitudinal profile morphology of the actively growing folds. Most old rivers show a concave profile, such that slope decreases in a downstream direction, adjusted to variations in lithology with distance downstream on various types of rock (see Appendix 1 catchment 2, 3, 16, 43, 44, 47, 48).

Finally, a comparison of maps compiled on the basis of photo interpretation and field data (Plates 1 and 2) show a close correlation with stream gradient observations as shown in Plate 4. The method presented here contributes to obtaining quantitative characteristics of late Quaternary deformation by allowing for the preliminary delimitation of areas subjected to different tectonic movements and by filtering out those anomalies related to other causes such as lithology. The former shorelines are elevated well above present-day sea level along the coastline of the eastern Waipara region, indicating both regional and differential uplift. These data were obtained on small drainage basins.

CHAPTER SIX

GEOMORPHOLOGICAL APPROACHES TO THE STUDY OF NEOTECTONICS

6.1 INTRODUCTION

In recent years there have emerged a number of new geomorphological approaches to the study of the effects of neotectonics of various types on alluvial rivers (Adams 1980b, Schumm et.al 1982, Ouchi 1985, Campbell & Yousif 1985, Bull & Knuepfer 1987, Harvey & Wells 1987).

The objective of this chapter is to discuss the use of geomorphic features to document the tectonic history of actively growing folds and faults. The concept of geomorphic thresholds is invoked to explain the interrelationship between uplift and the landforms developed during the evolution of the fluvial drainage system.

In order to provide the background necessary for a study of the possible effects of neotectonics on the evolution of the fluvial drainage system some basic concepts in fluvial geomorphology are reviewed.

6.1.1 THE CONCEPT OF FLUVIAL PROCESSES

"Everything about the fluvial system reflects instability and variability in process and rate" (Schumm 1981).

A fundamental fluvial-process concept is base-level adjustment and entropy. Base-level is a datum below which streams, at a specific time, can do no work. Physically, this datum may be represented by features such as strath surfaces, pediment terraces or erosion glacis.

Entropy in a fluvial system refers to a tendency for achieving a least-work/maximum probability energy grade (Shannon & Weaver 1949, Pelto 1954, Miller & Kahn 1962, Leopold & Longbein 1962, Thornes & Brunsden 1977, Bull 1979) and this is reflected in channel hydraulic geometry and basin geometry.

Several prominent terrace levels in the lower Waipara valley and other subhorizontal alluvial surfaces within large and small drainage basins testify all three modes of stream operation (i.e. aggradation, degradation and equilibrium or grade) throughout the late Quaternary. Equilibrium conditions are associated with stable longitudinal profiles, and aggradation and degradation are disequilibrium modes of operation that occur where insufficient time has elapsed since tectonic and/or eustatic and/or climatic perturbations for interactions between variables to result in equilibrium conditions. In response to one of the above constraints streams can adjust internally and externally. The response of the fluvial system to changes in these controls is complex involving both erosion and deposition and to subtle changes in bedform and valley profile within these regimes. Degradation and aggradation modes of operation of streams and changes in bedform are separated by the threshold of critical power (Bull 1979).

6.1.2 EFFECT OF ACTIVE TECTONIC DEFORMATION ON ALLUVIAL RIVERS

Rivers respond to valley-slope deformation caused by active tectonics in various ways depending on the rate and amount of surficial deformation and the type of river. Unfortunately for tectonic studies they adjust to changes of hydrology, lithology, sediment load and type, human intervention as well Schumm 1977 & 1981,

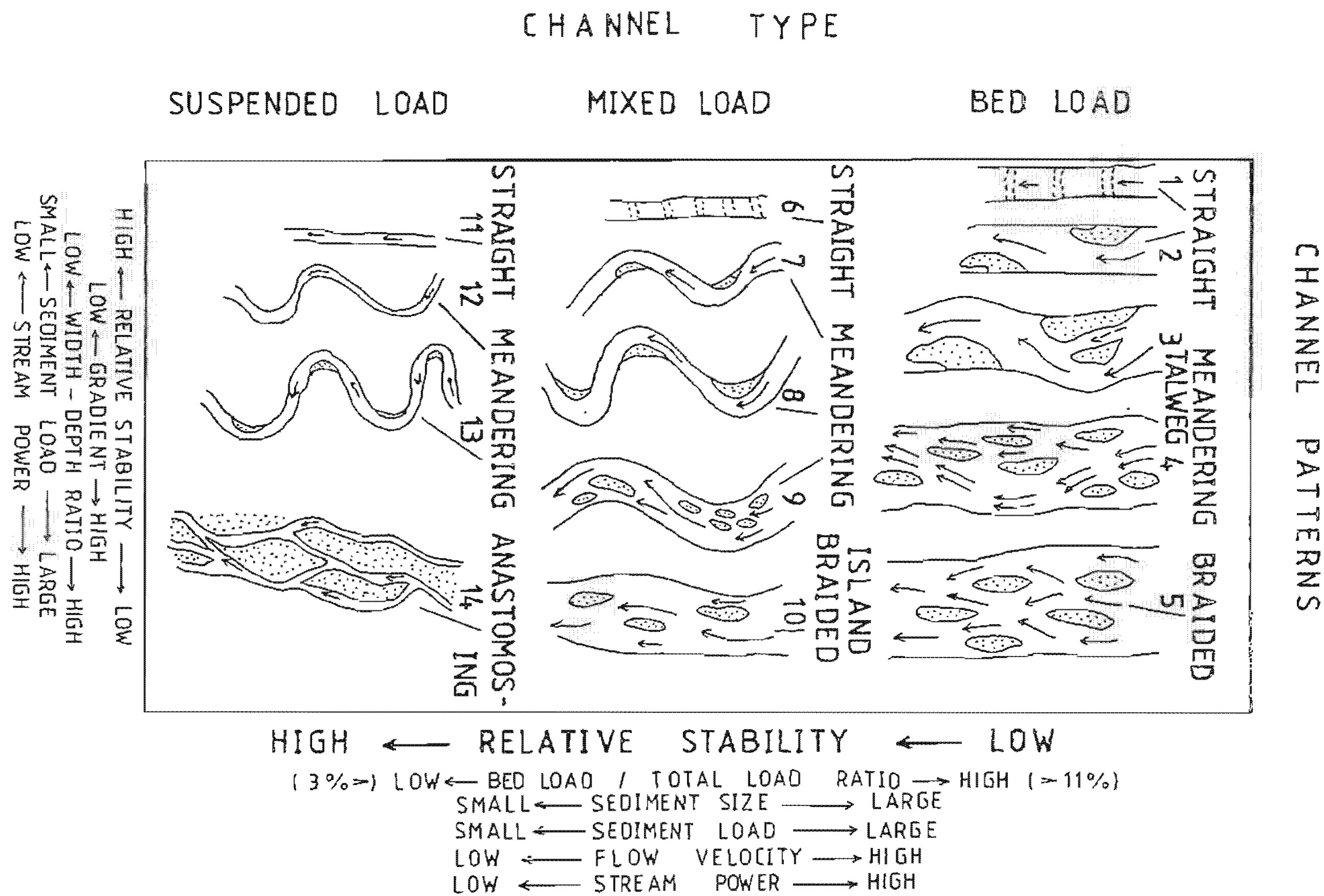


Figure 6.1 The range of alluvial channel patterns (from Schumm 1981).

Cooke & Reeves 1976, Ouchi 1985.

Schumm (1981) developed a classification of alluvial-channel patterns, applying the concept of channel-pattern change with valley slope to the classification of alluvial-channel pattern with sediment load types (Fig. 6.1). The patterns are divided into three major channel types. Decrease in gradient can cause a pattern shift from right to left within a major channel type. Increase in gradient has the opposite effect, possibly causing a shift from left to right. Although it is still hypothetical, this classification is pertinent to this study, because it appeared to be compatible with the observations reported here. Using this classification, hypothetical models of alluvial-river response to active tectonic movement (anticlinal uplift, synclinal subsidence and reverse faulting) was developed for a braided (bed load) river (Fig. 6.2). The response will differ, however, depending on the proximity of the channel characteristics to a threshold value.

In Figure 6.2 the decreased gradient upstream of the axis of an anticline is similar to the reduced slope downstream of the axis of a syncline and upstream of a tilted reverse fault block. It is important to realize that channels that lie near a pattern threshold may change their characteristics dramatically with only a slight change in the controlling variable. For example, some rivers that are meandering and that are near pattern thresholds become braided with only a small addition of bed load (Schumm & Khan 1972, Schumm 1979).

However, the clearest evidence for tectonic effects on rivers will be anomalous reaches showing significant changes of pattern, gradient, and valley morphology that cannot be atttributed to other causes (Yousif 1986). The deformation of valley slope (Fig. 6.2),

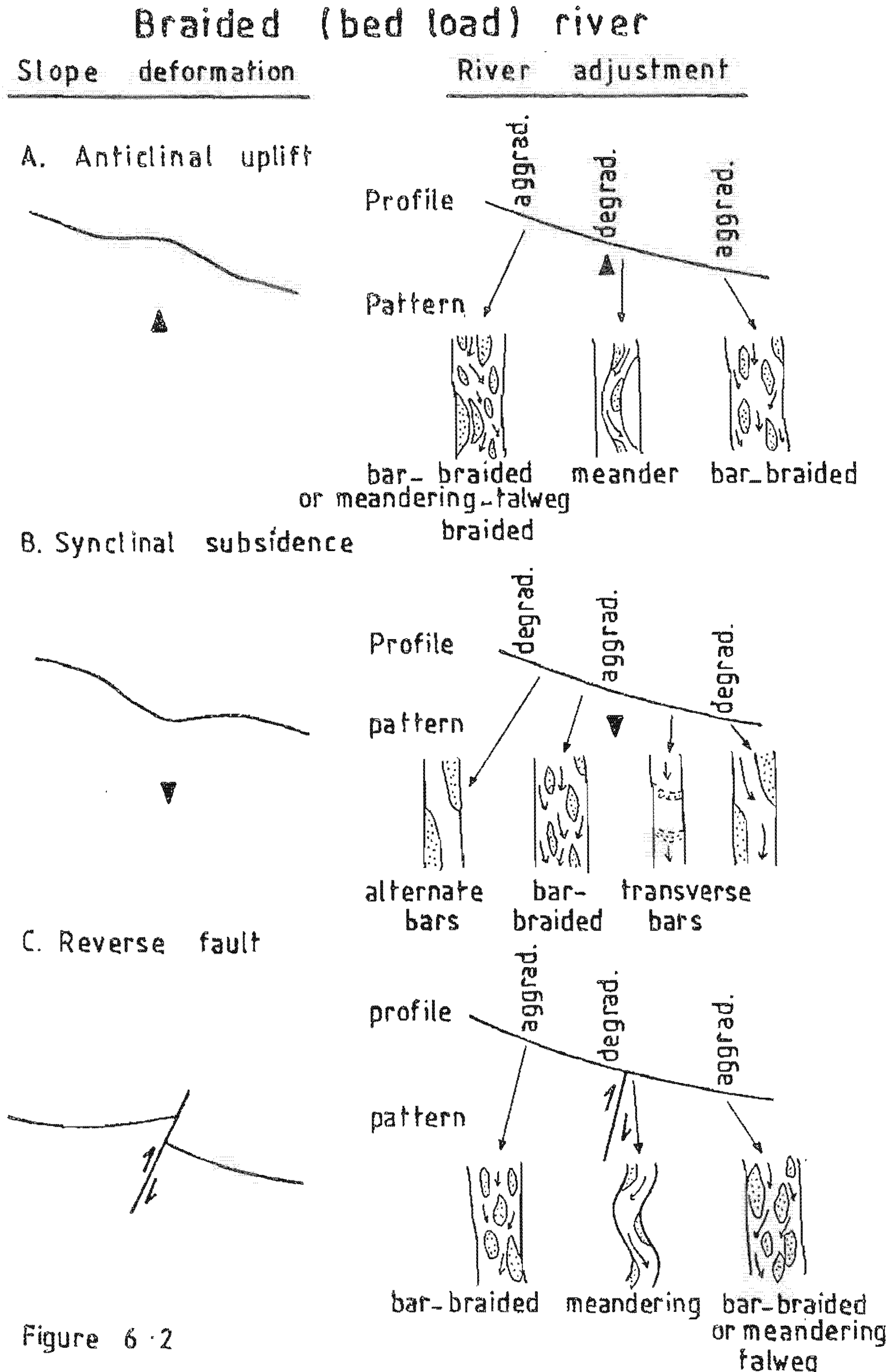


Figure 6·2

Adjustment of a braided river to (a) anticlinal uplift, (b) synclinal subsidence and (c) reverse fault across it (modified after Ouchi 1985).

the slope of the surface on which the channel is formed, will inevitably change channel gradient, which is a dependent variable determined by water and sediment discharge and by sediment size. This change of channel gradient will upset the equilibrium between channel slope and hydraulic properties of the stream.

Evidence for continuing effects of deformation on channel morphology is demonstrated by further investigation such as bank height, gradient, channel aggradation or degradation, and change in channel sinuosity. These will be discussed in detail in relating subsections of this study.

6.2 CASE STUDIES

This chapter involves seven case studies chosen for contrast in the dominant modes of deformation, catchment size, and erosive power of the principal river. The first six case studies are located within the study area, the seventh example, the Rakaia River, to the south of the study area.

1. The Waipara River
2. The Carrington Creek
3. The Yellow Rose Creek
4. The Kate Creek
5. The Teviotdale River
6. The Motunau River
7. The Rakaia River

The distribution of known and probable fluvial surfaces was determined by aerial photograph examination of the field areas, and plotted on 1:10,000 and 1:5,000 scale maps. Two precise levelling surveys of the lower Waipara River and Carrington Creek were made.

6.3 THE WAIPARA RIVER

6.3.1 SETTING

The Waipara River and its tributaries forms the major drainage basin in the study area. The Waipara River rises in the eastern foothills of the Southern Alps, at c. 840 m elevation, flows eastward across the Canterbury Surface to Waipara, then southeast to the sea. The main channel length is about 63 km, draining a total area of about 961 square kilometers. This study includes only the lower section downstream of the junction with Weka Creek (i.e the lower 15 km of the river valley Fig. 6.3).

The behaviour of the Waipara River at the point where it crosses the axis of the Cass Anticline demonstrates anomalies in the valley and channel characteristics which provide three pricipal lines of evidence in support of a history of antecedence across an actively developing fold.

1. From the headwaters to the sea, the Waipara River crosses two structural barriers in antecedent gorges and the bed form alternates between braided reaches of aggraded flood plain and deeply incised channels in bedrock. In the lower gorge (Stereomodel 17), the transition into the upstream (west dipping) flank of the Cass Anticline is marked by an abrupt change to an anomalous incised meander pattern which straightens again downstream of the axial trace. A marked change in sinuosity across the axial trace of an active fold is a phenomemon noted by several authors but with some differences which will be discussed shortly.
2. The lower Waipara gorge is a portion of the valley with

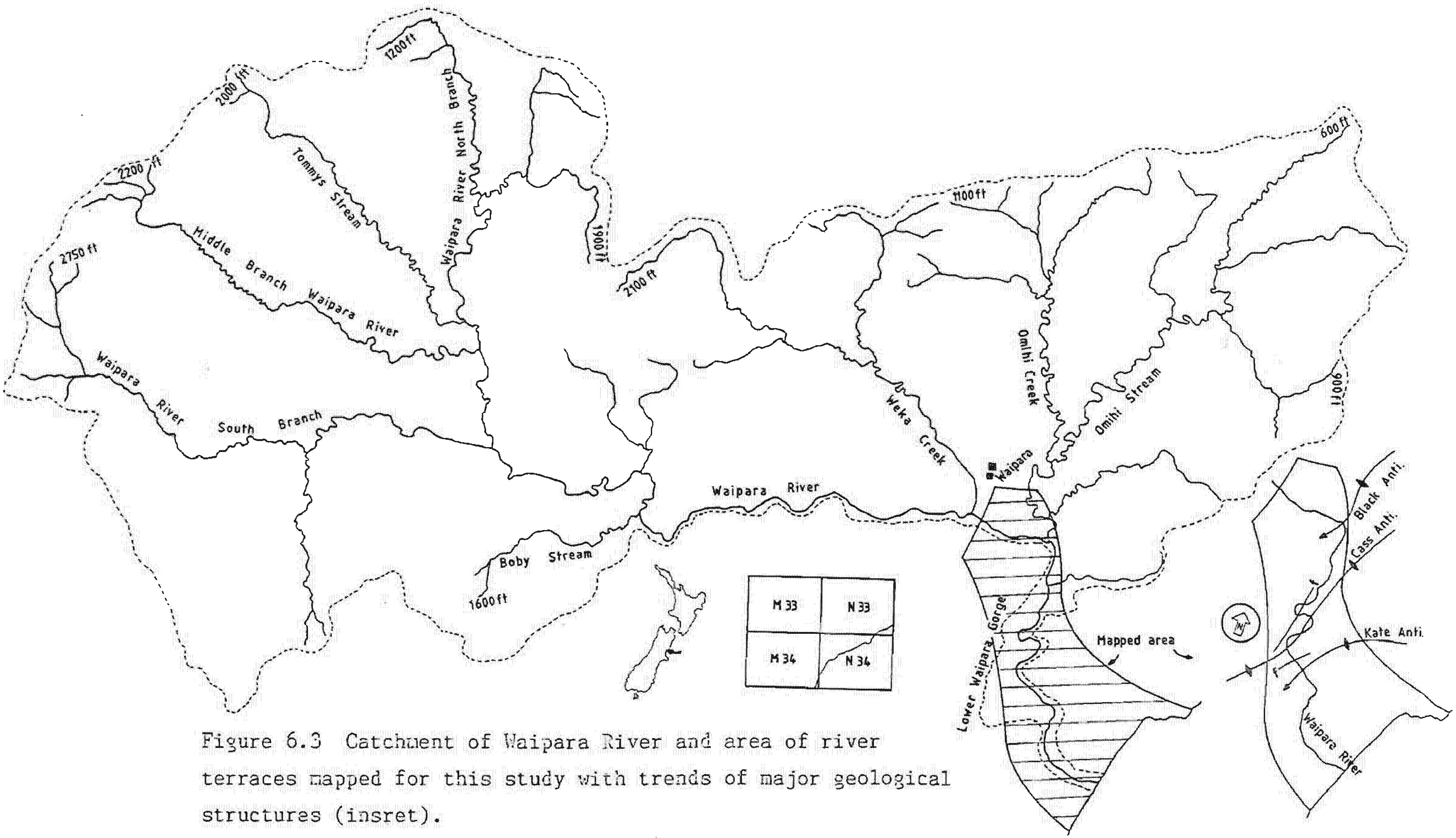


Figure 6.3 Catchment of Waipara River and area of river terraces mapped for this study with trends of major geological structures (insret).

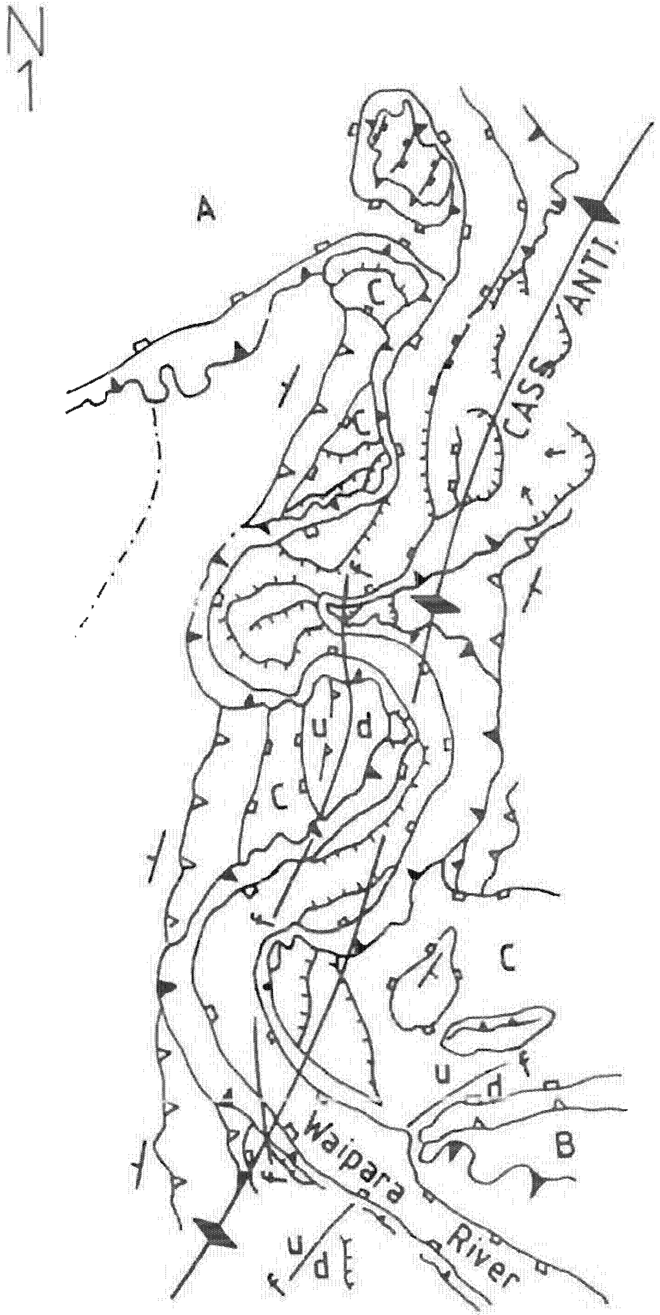
almost vertical bedrock walls rising 50 to 80 m above the present river. The narrow gorge, only about 150 m wide, represent a distinct topographic departure from the broader Waipara valley both upstream and downstream. Everywhere in the gorge, the river is entrenched deeply in the bedrock, and whatever alluvium is present occurs as bed load, river gravel capping small terrace remnants, or as alluvial fan debris.

3. Realignment of the river course and the profiles of uplifted river terraces provide a good example of the nature of channel response to uplift in this area.

No previous detailed studies have been made of the lower Waipara River terraces, although Wilson (1963) mapped a series of terraces on both sides of the river without attempting cross river correlation. Gregg (1950) was the first to mention the importance of employing accurate surveying methods to obtain data on the heights and gradients of terraces which would undoubtedly do much to elucidate the story of the evolution of the present course of the river. Terrace remnants underlain by alluvial deposits occur at a variety of elevations along the study area; these remnants are found from the level of the modern flood plain to the level of the upla nd Bobs Flat marine terrace of probably the Last Interglacial stage (Plate 7). Terrace remnants less than 30 m above river level are scarce in the meandering valley, but they are widespread across the Canterbury Surface just upstream and downstream from the gorge, where they occur on both sides of the river.

Stereomodel 17

Aerial view of the Lower Waipara Gorge showing the course of the Waipara River cutting through the southern end of the Cass Anticline. Of particular interest is the antecedent gorge. Note the ancestral (area C) and modern river terraces, cross-valley morphology, fault traces, and upper and lower aggradation surfaces (area A and B respectively).


N
A
C
CASS ANTI.
u
d
f
Waipara River
B

6.3.2 GEOMORPHIC FEATURES

Based on observation of individual geomorphic units and landforms from dynamic, tectonic, and morphogenetic aspects, the area was divided into eight geomorphic surfaces (Plate 7).

A. Last Interglacial (Bobs Flat) Marine Terraces

The highest and oldest marine terraces preserved on the coast exists along the southern limb of the Kate Anticline in the region north of the Ancestral Waipara River mouth (Plate 7). They are deeply dissected by small coastal streams.

Based on their heights and geomorphic positions, the Bobs Flat coastal plain of Carr (1970), can be subdivided into three levels. The times of formation of the three marine terraces are tentatively assigned to the three levels of the Last Interglacial age (oxygen isotope Stage 5, c.80,000-130,000 years B.P.), on the evidence previously discussed in Chapter 5, section 5.7. The former shoreline representing the maximum height of sea level attained during terrace formation is buried beneath slopewash deposits at the base of the riser. No marine deposits were seen at M4 and M3 but they are regarded as marine in origin (see Chapter 2 section 2.6.3). The type locality for the lower marine terrace (M2) is about 106m above sea level (S68/165054). The terrace deposits consist of fine beach gravel, well weathered, and cemented by calcite into a hard mass (Fig. 6.4). Global sea level at 80 Ky was (-13) m lower than that at present, hence the average rate of uplift of the terrace since that time was about (106-(-13)/80=1.48 m/1000 years or 1.48 mm/yr.

Significant tectonic tilting of this unit is indicated by altimetric data, but the altimetric data is of insufficient quality to allow the degree of tilt to be determined accurately (Carr 1970).

Field evidence indicates that the ancestral Waipara River gravels were accumulating contemporaneously alongside the beach deposits of the lower marine terrace (M2). A well preserved yellow weathered alluvial greywacke gravel, estimated at 3 m thick, unconformably overlies the beach deposits (S68/161052). The pebbles are mostly subrounded and up to 7 cm in diameter, the whole deposit being similar to the younger river gravel of the Teviotdale Gravel. The absence of greywacke in the drainage basins of local streams suggests an external source for this river gravel and that the surface of the west part of the lower marine terrace is a river aggradation terrace that accumulated during the last high sea level of the last interglacial at c.80,000 years B.P.

B. Teviotdale Surface

The Teviotdale Surface is the oldest aggradation surface to be found along this stretch of the river. Gravels underlying the terrace surface rest with marked unconformity upon older rocks (Fig. 6.5a) and consist of yellow-brown, very poorly sorted pebble gravel with rare pods or lenses of finer material, mostly very fine sand and silt. Pebbles are composed almost entirely of greywacke, and are subangular to rounded (Fig. 6.5b). The gravels have been deposited during a period of fluviatile aggradation of the ancestral Waipara River. The unconformity at the base of the gravel section is regarded as a surface of stream planation (i.e an ancestral strath). This aggradation phase is presumed to have formed during 70-75 Ky B.P.

This geomorphic unit occupies three isolated areas. The first is located on the south-eastern limb of the Cass Anticline, which extends

Figure 6.4

View looking southwest along the southern limb of the Kate Anticline showing Bobs Flat marine terraces.

Figure 6.5a

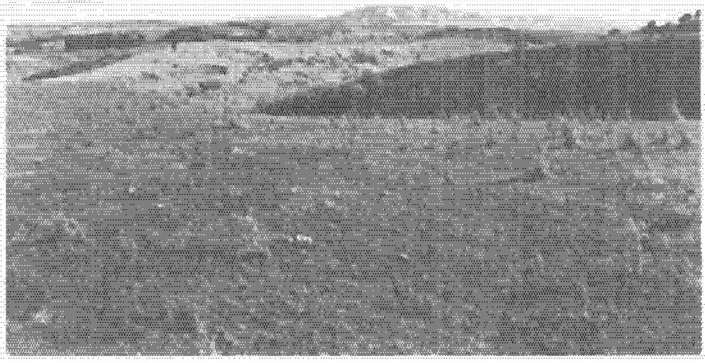
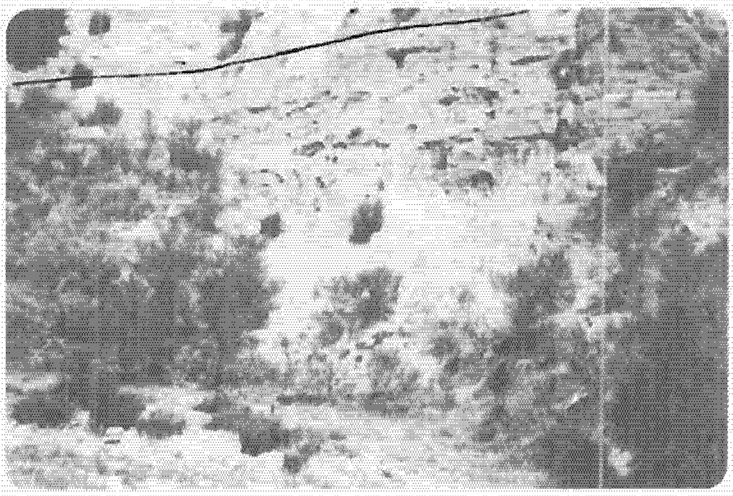
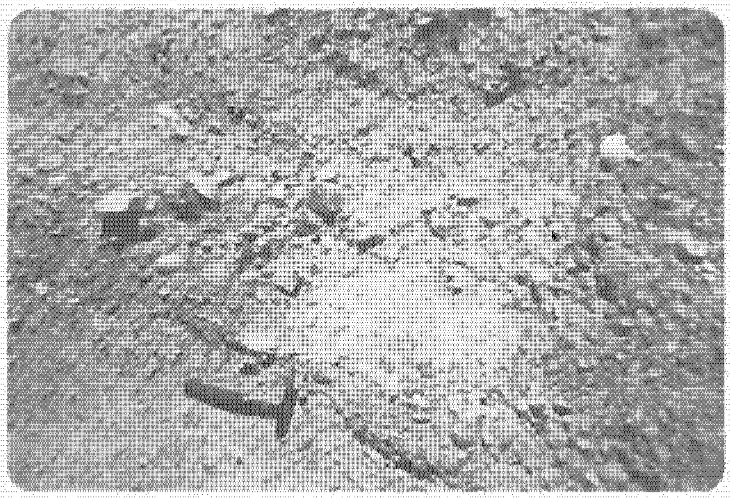
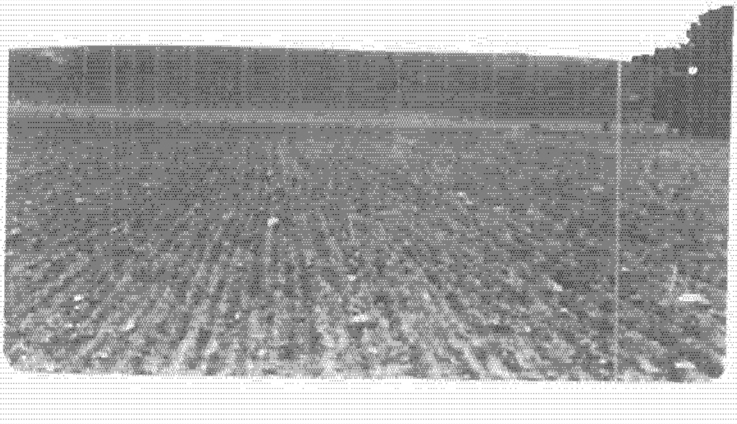
Waipara River bank exposure downstream of Greenwood's Bridge, showing angular unconformity between Teviotdale Gravel and Greenwood Formation (arrowed).

Figure 6.5b

Close up of the Teviotdale Gravel. Note poor sorting and high percentage of matrix.

Figure 6.5c

The Mound, a flat-topped remnant of the Teviotdale Surface dipping gently south-east. Canterbury Surface in foreground.

for about 2.4 km from just below the lower Waipara Gorge towards the sea. Here a number of small streams transect the surface and a remnant of this surface was also found along the north side of the ancestral Waipara valley below Bobs Flat marine terrace (Plate 7). The second area is generally aligned along the north-western limb of the Cass Anticline for about 1.0 km west of the Waipara River towards the main north highway. Finally there is the Mound (S68/095117) about a 1.5 km west of the Waipara Bridge and about c. 30 m higher than the Canterbury Surface (Fig. 6.5c).

C. Last Glacial (Tiromoana) Marine Terrace

The next youngest marine surface occupies a narrow uplifted coastal plain north of the Waipara River mouth, extending for about 1.5 km towards Teviotdale River. At the inland margin of the unit is a marine cliff cut into the fluvial aggradation gravels (Teviotdale Gravel). The cliffing is estimated to be formed as the result of 60 Ky interstadial transgression. The age of 20 Ky Kawakawa Tephra beneath this terrace at (S68/156038) is of certain significance in relation to the estimate age of the above transgression, on the evidence previously discussed in Chapter 5, section 5.7. The average rate of uplift of the coast since that time was about (46- (-21) /60 = 1.11 m/1000 years or 1.11 mm/yr.

The present distribution of this terrace has been influenced by active rill, gully and coastal erosion, which has removed all trace of the terrace surface from some areas. Erosion by the ancestral Waipara River has altered the aspect of the riser and the extent of the terrace near the river mouth and has reduced the thickness of Tiromoana Formation from 21 m to 4.5 m., and subsequent aggradation has left a terrace cover of 8 m of river gravel and silt (Fig. 6.6). Here the basal member of the Tiromoana Formation rests upon a surface planed on the Tokama Siltstone.

D. River Terraces less than 20,000 yrs B.P.

The following degradational phase produced at least four terraces here termed the ancestral Waipara terraces. Terraces within this age group form the majority of those preserved in the Teviotdale valley (i.e. ancestral Waipara valley) and are identified by the absence of Kawakawa Tephra in coverbeds and by height relationships. Near the Teviotdale Homestead (S68/145064), the ancestral Waipara River once flowed down the valley now occupied by Teviotdale River. It has a broad U-shaped character and four distinct terrace levels (Fig. 6.7a). At (S68/148055), terrace no. 3 comprises a series of discontinuous erosional escarpments cut in to the Greenwoood Formation. The lower terrace is a low-lying belt of land between the mouth of the Teviotdale River and the beginning of the gorge near the Teviotdale Homestead. Plate 7 shows the distribution of river terraces in this age category within the field area.

Flights of narrow discontinuous tilted erosional terraces, a broad U-shaped valley, air gap, and strath surfaces above the modern floodplain have been distinguished throughout the gorge, bordering the meandering course of the lower Waipara River (Fig. 6.7b). Unmatched terraces produced by meander slip-off are formed by the lateral migration of the continuously entrenching river (Plate 7).

E. Canterbury Surface

The lowest and youngest aggradation surface found along the

Figure 6.6

Unconformity between beach deposits and overlying aggradation gravels at sea cliff near ancestral Waipara River mouth.

Figure 6.7a

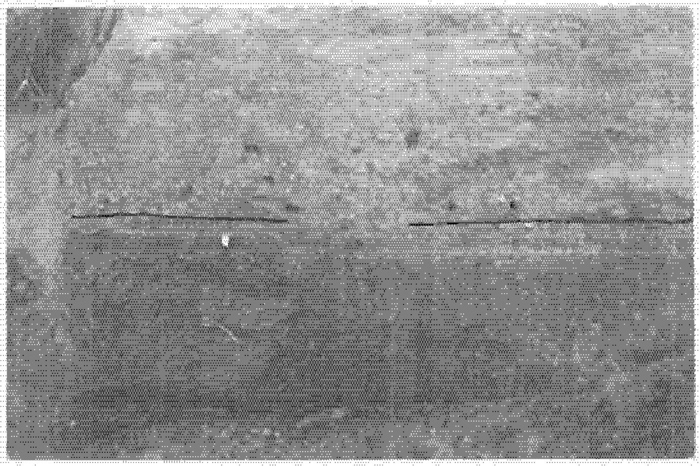
Erosional terraces along the ancestral Waipara valley viewed from Teviotdale Homestead. View looking east.

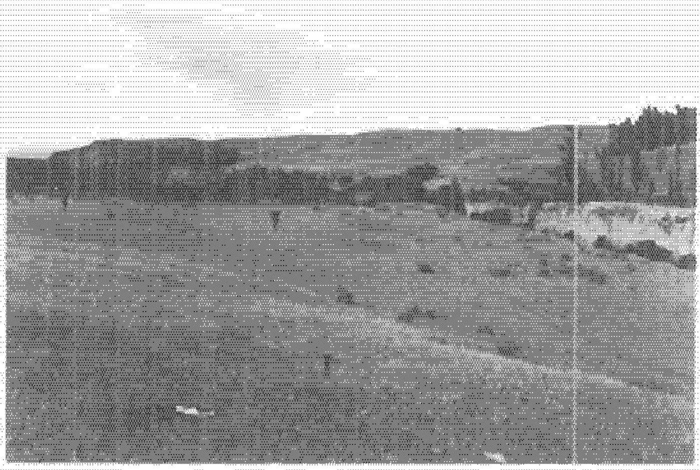
Figure 6.7b

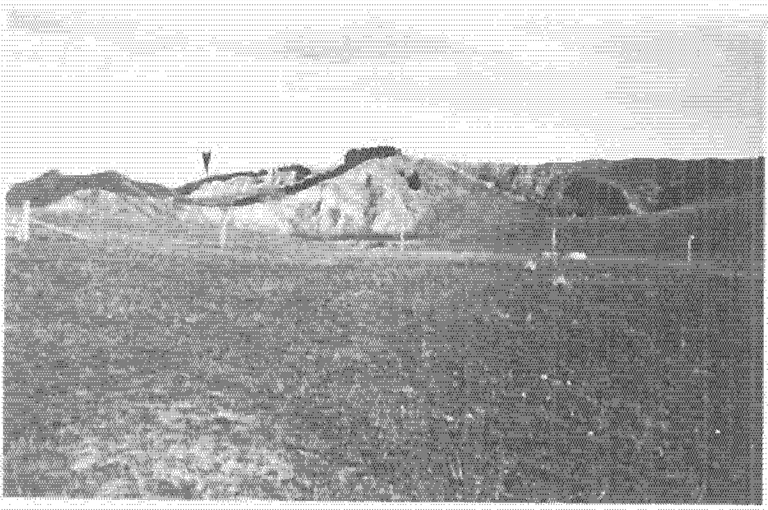
Landforms along the lower Waipara gorge showing ancestral valley cross-section and tilted erosional terraces in foreground.

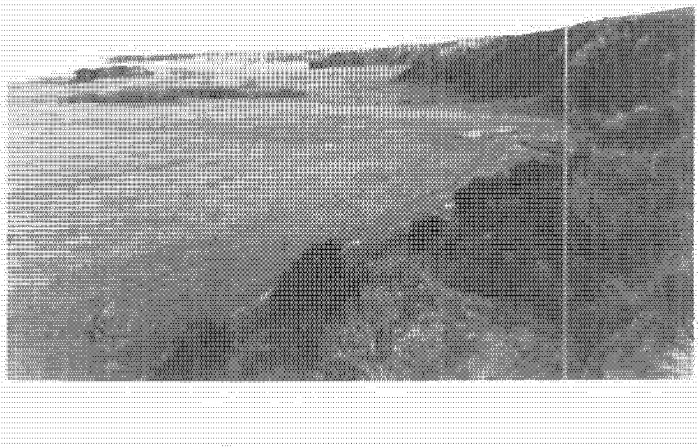
Figure 6.8

View facing southwest from Teviotdale River mouth, showing Amberley coastal plain in foreground and postglacial marine cliff to the right.









lower Waipara River is much wider in extent than the older terraces (Plate 7). The Canterbury Gravels comprise the whole of the material deposited by aggrading rivers, culminating in the formation of the surface of the Canterbury Plains (Wilson 1955).

F. River Terraces less than 10,000 yrs B.P.

A sequence of ten erosional terraces has been identified below the level of the Canterbury Surface, but predating the Amberley coastal plain. The greatest number of river terraces is preserved near the Waipara Bridge and river mouth, where 10 terraces are preserved from 73 meters and 38 meters above sea level (present flood plain c. 55 m and 8 m a.s.l. respectively). Throughout the gorge, the Waipara River was entrenched into the bedrock, and widened its channel by lateral cutting. Gravels underlying the terrace surface are blue-grey, fresh, medium, poorly sorted gravels, with only a small proportion of sands and silts that form the matrix. The pebbles are rounded, and principally greywacke, but there is a higher proportion of Tertiary pebbles than in the Canterbury and Teviotdale Surfaces.

G. Amberley Coastal Plain

This unit occupies a gently sloping surface of land lying between the present shoreline and the postglacial marine cliff (Fig. 6.8). On aerial photographs, two coastal plains can be distinguished: 1) a coastal plain with elongated, subparallel ridges, with inter ridge depressions that are frequently wet, and 2) a coastal plain with smooth, cultivated beach ridges (Stereomodel 10). Gregg's (1964) map shows this coastal plain and its extensions as Holocene.

H. Flood Plain and Valley Filling

The most obvious area in this category is the modern river bed and its surrounding flood plain. Two flood plains are distinguished: 1) active flood plain with abandoned channels and sand bars, and 2) inactive flood plain such as that adjacent to the Waipara River (Stereomodel 10). Lithologies in the present Waipara River channel gravel are greywacke derived largely from the headwaters in the foot hills of the southern Alps, and Tertiary rock derived from outcrops in the upper and middle sections of the river. The amount of Tertiary rocks present in the river is dependent upon proximity of tributaries supplying these rocks and erosion of Tertiary bedrock underlying the present river bed.

6.3.3 GEOLOGICAL HISTORY OF THE LOWER WAIPARA RIVER

The major physiographic feature of the lower Waipara gorge is the gorge itself. The peculiar situation of the gorge has attracted the attention of numerous geologists.

Three explanations have been put forward concerning the history of the lower Waipara gorge .

1. The hypothesis that the gorge is antecedent (Cotton 1913; Speight 1915):

This hypothesis requires that Waipara River was in existence before tectonic movements took place. During uplift, the erosive power of the river was sufficient to incise, and the river continued its course, so cutting the lower Waipara gorge.

2. The hypothesis that the gorge is superimposed (Hutton 1905; Speight 1912; Jobberns 1937a,b ; Gregg 1950):

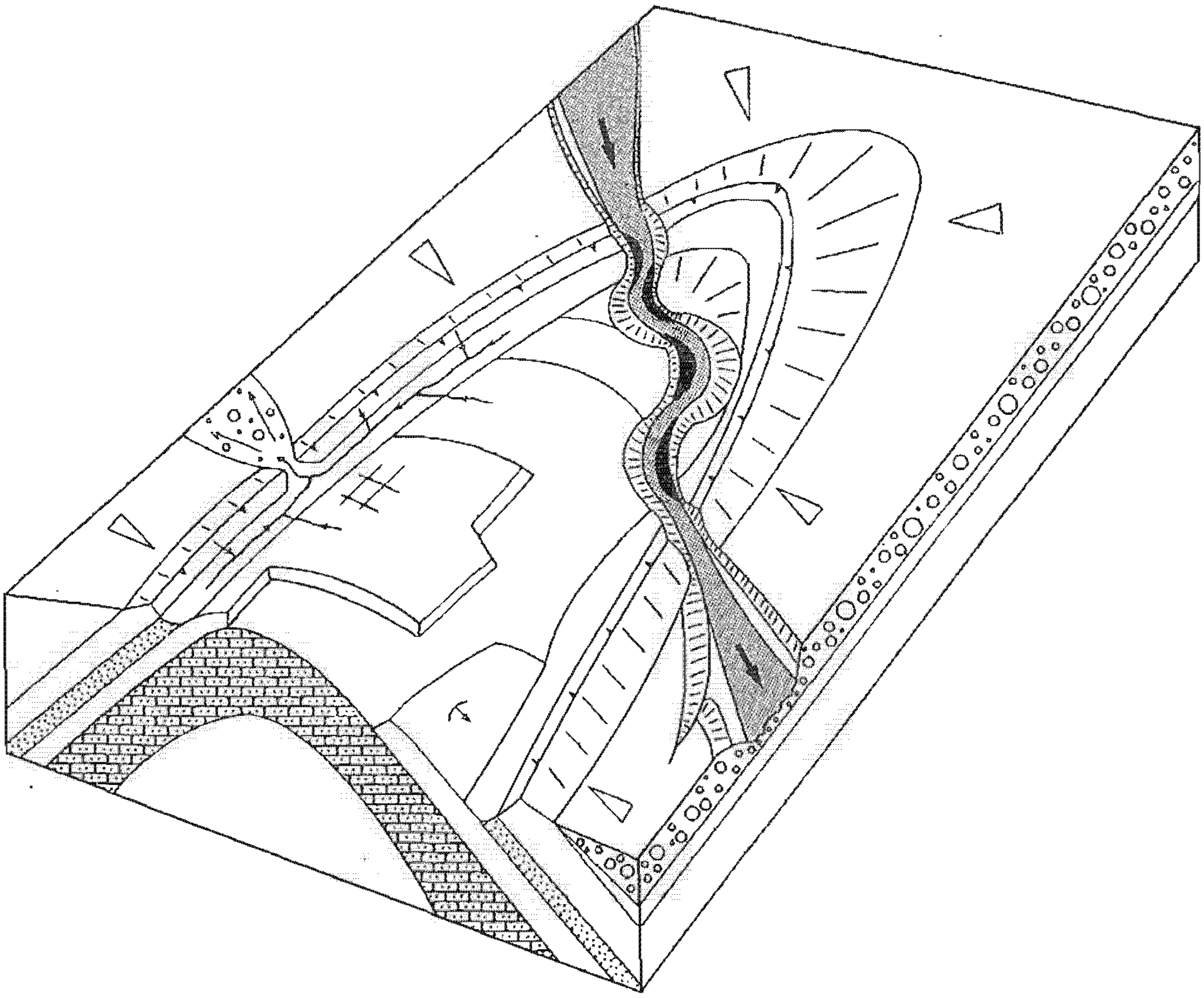
A superimposed valley is formed where previously folded or tilted rocks are buried by younger unconformable sediments, on which a consequent drainage then forms. This drainage cuts vertically downward into the buried structures and becomes incised into them.

3. The hypothesis that Waipara River was captured by a small headward-eroding coastal stream (Speight 1912).

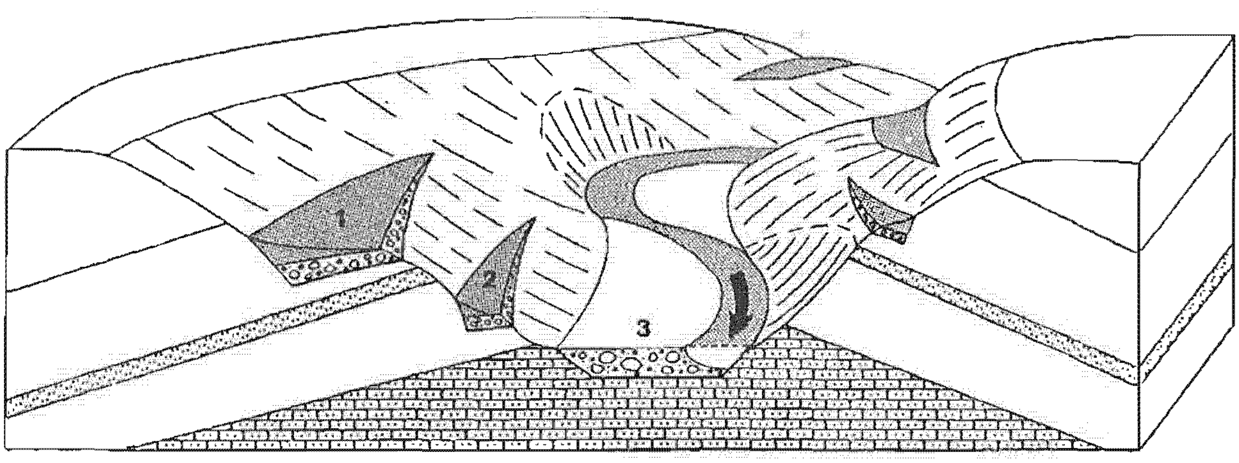
6.3.4 ANTECEDENT HYPOTHESIS

I agree substantially with the antecedent origin for the lower Waipara River (Fig. 6.9a). The manner in which streams are influenced by uplift is related to many factors, including discharge, gradient, topography, and lithology. Experimental studies and field observations confirm that a change of valley floor slope will cause a change of channel morphology (Schumm & Khan 1972, Ouchi 1985). For example, upstream of the axis of uplift there should be a decrease of the valley floor slope due to backtilting, whereas downstream of the axis of uplift there should be an increase in valley floor slope. A stream crossing the uplift should be affected by such slope changes to produce a reduction of competence and stream power in the area upslope of the axis and an increase of competence and stream power downslope of the axis. Adams (1980b), King and Stein (1983) suggested that rivers are responding to the valley slope changes by changing their sinuosity in order to maintain equilibrium slope. Schumm et al. (1982) noted that tilting of alluvial terraces in a valley is the most convincing evidence of deformation. The oldest terrace will be the most deformed if deformation has persisted over a period of time.

The ratio of stream power to uplift in the study area has been such that during the evolution of the drainage at any period, some



A. Conceptual diagram of an antecedent gorge



B. Conceptual diagram of tilted erosional terraces with meandering river

Figure 6.9

Block diagrams of the antecedent model of the lower Waipara River.

reaches of most streams have been close to the critical-power threshold. There is considerable evidence supporting the view that major and small streams in the study area have had erosive capacity in excess of that needed to keep pace with lower base levels, during part or all of their history (Campbell & Yousif 1985). They are thus capable of crossing one or more thresholds, as can be inferred from the following observations:

1. The cross-valley morphology of the floor of the gorge is flat carrying only a bedrock veneer of gravels (Fig. 6.10a), and it is evident that lateral planation is occurring in the bedrock to form a new strath surface at the present time (Fig. 6.11). Data taken from the new Greenwood's Bridge.
2. Remnants of ancestral straths are now preserved under terrace gravel above the modern flood plain (Figs. 6.10 b,c,&d).
3. A change in valley slope can significantly change channel patterns and sinuosity. A dramatic change of pattern, from braided through meandering to braided, of Waipara River in the lower Waipara area occurred both spacially along the length of the river at points showing clear relationship to dip directions in the fold limbs and with time as evidenced in the gorge cross-profile.

Where there is an actively growing anticline, oversteepened gradients of old fluvial deposits are to be expected on the downstream side of the uplift. The gradient on the upstream side would be reversed. There are no theoretical reasons on geomechanical grounds for expecting uplift rates of growing anticlines to be uniform over the entire history of fold growth. Furthermore, periods of rapid

Figure 6.10a

Landforms along the lower Waipara gorge showing modern flood plain, broad U-shaped valley and lower erosional terraces.

Figure 6.10b

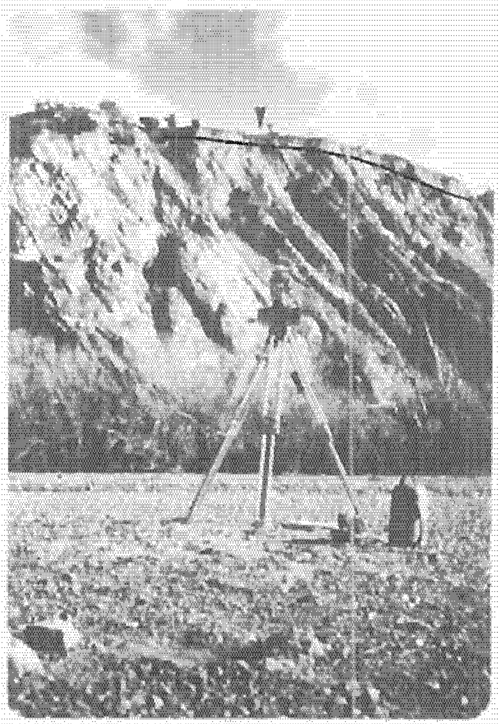
Waipara River bank exposure, at S68/140082, showing ancestral gravels resting unconformably on highly tilted Greenwood Formation.

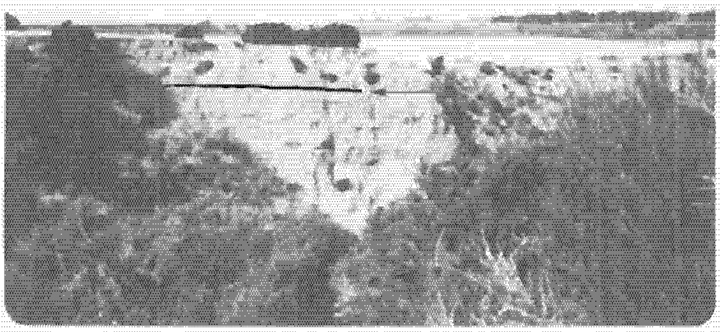
Figure 6.10c

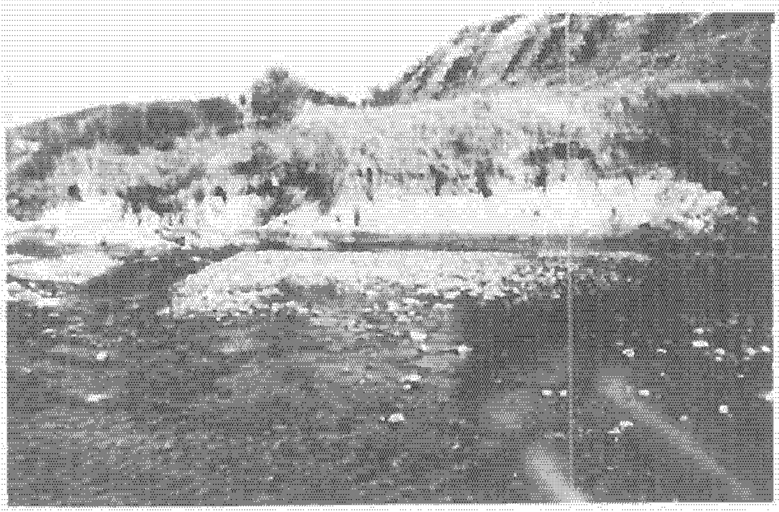
Waipara River bank exposure, 3 km upstream from mouth, at S68/140040, showing near-horizontal ancestral gravels resting unconformably on tilted Kowai Gravels.

Figure 6.10d

Waipara River bank exposure in the middle gorge at S68/134070, showing lateral planation is occurring in bedrock to form a new strath surface at the present time.







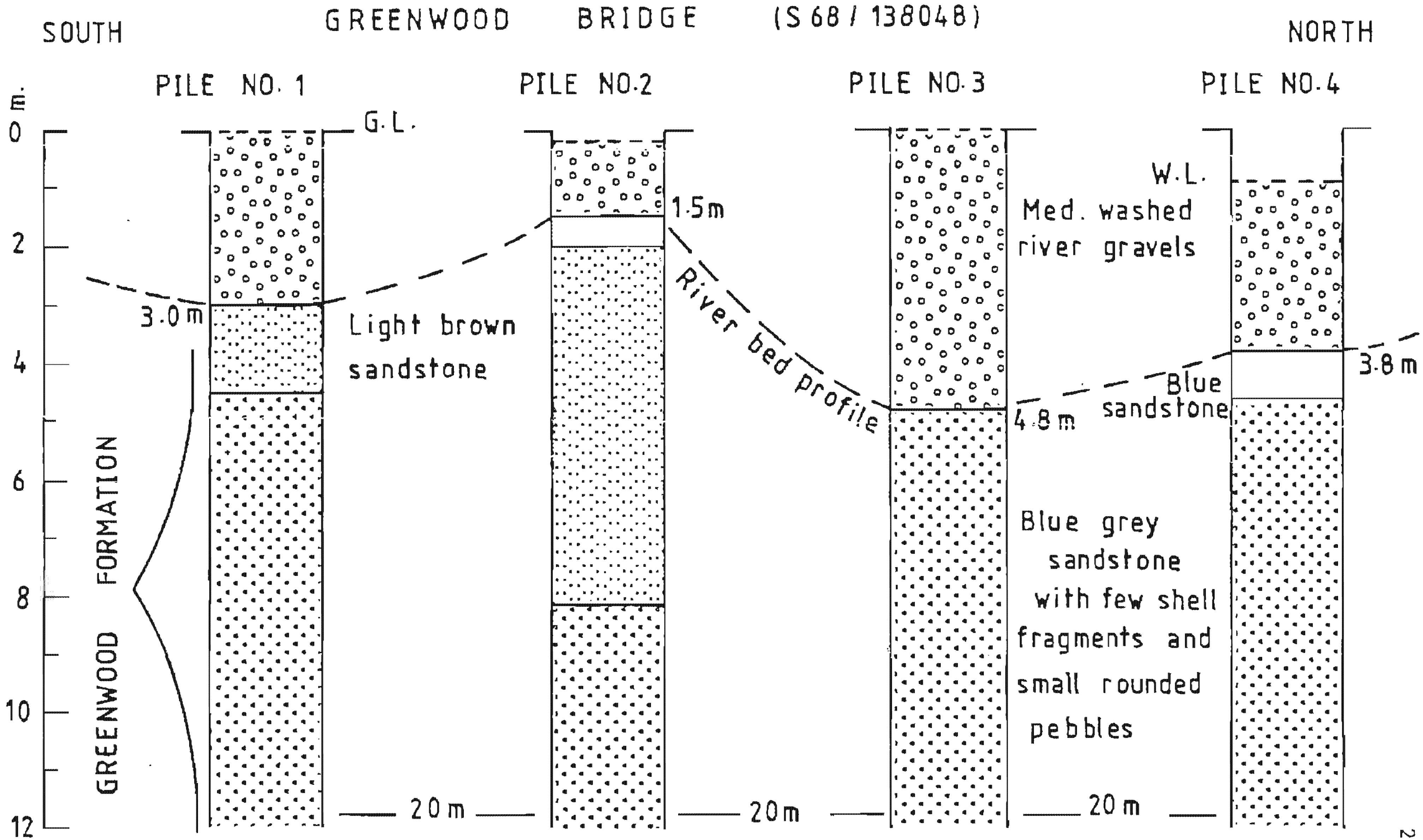


Figure 6.11
Cross-section of the lower Waipara River valley.
(Greenwood Bridge bore hole data).

uplift may be separated by periods of minimal tectonism, particularly if folding is related to intermittent fault movement at depth, as is likely to be so here. Straths formed during periods of tectonic quiet will become terraces with the onset of the next period of accelerated uplift, which steepens the slope of the active stream channel (Fig. 6.9b). Thus strath and cut stream terraces represent time lines along valleys because they are formed during periods of equilibrium or threshold conditions (Bull 1979). During this equilibrium phase, the river is presumed to be adjusted with regard to its load and discharge conditions and, therefore, migrates laterally across a stabilized valley bottom. As lateral migration proceeds, the river truncates underlying bedrock while simultaneously depositing a thin veneer of point bar gravel on top of the bevelled bedrock surface. Thus, an erosional terrace should be underlain by a thin sheet of channel sediment spread evenly across a flat, laterally eroded rock surface. As shown on Fig. 6.9b, therefore, the oldest strath deposits such as those in (1) would be deformed the most. The gradient of strath deposits such as those in (2) would be only slightly increased, and the modern strath deposits would be unaffected. In the uplifted area, the valley is narrow and steepsided; upstream and downstream the valley can be broad and flat. In this way topographic anomalies result.

6.3.5 SINUOSITY

The sinuosity of a stream over any particular reach is defined as the ratio between the channel thalweg length and the corresponding valley length (Schumm 1977). The sinuosity values of the modern Waipara River are shown in (Fig. 6.12).

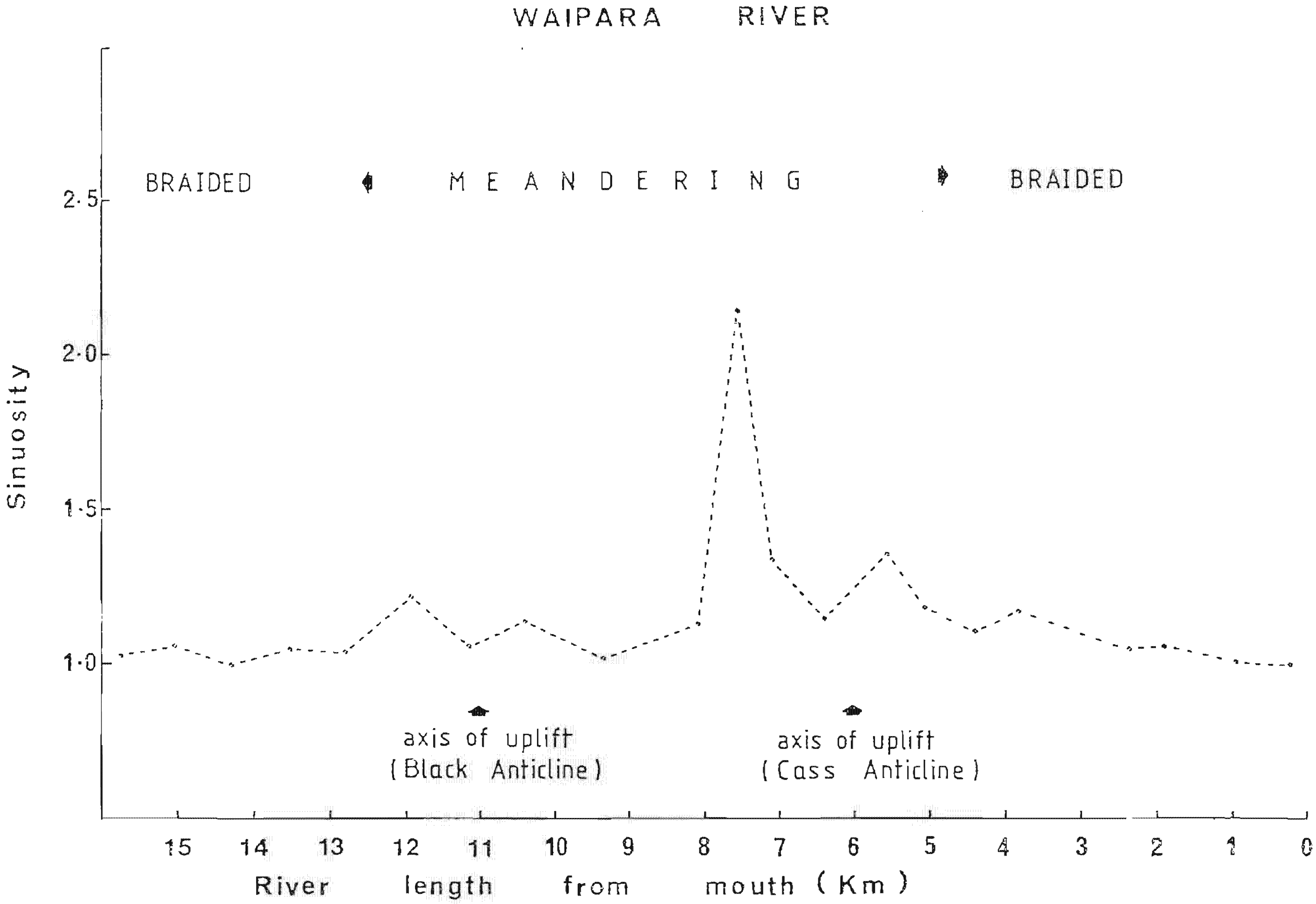


Figure 6.12 Realtionship between river sinuosity, axis of uplift and distance.

From this plot, the upvalley side of the uplift axis has higher values of sinuosity than immediately downvalley of the axis. These data show that variations in the valley floor slope have influenced sinuosity, as the river adjusts its pattern to maintain a constant gradient over the changing valley floor slope. This change in sinuosity of antecedent streams is well documented from several described examples and from flume experiments (Schumm 1977, 1981; Ouchi 1985), but the Waipara River appears to represent a system which has evolved further and probably reflects more complications in the deformation.

6.3.6 GEOMORPHIC EVIDENCE OF DEFORMATION

A. <u>Methods</u>

In order to identify displacement of a fluvial terrace by earth movements, a detailed survey of both strath terraces in the antecedent gorge and surface terraces in the lower Waipara area was made (See Appendix 2). The altitudinal relationships of the present channel and flood plain of the lower Waipara River to the adjacent terrace fragments have been projected onto a two-dimensional profile with height plotted against distance on a simple arithmetic scale (Plate 8). The following procedures have been adopted:

1) The riverbed profile has been calculated through the main centreline trend of the present Waipara River along the axis of the valley. Thus, the river centreline stations reflect the gradient of the flood plain. Between stations (generally spaced 200 - 300 m apart) the gradients of both flood plain and adjacent terraces and strath surfaces are projected onto a straight profile. Thus the longitudinal profiles shown in Plate 8 are not projected onto a single

plane surface of uniform bearing, but follow the line of survey stations shown on the geomorphic map, and reflect as clearly as possible the true down valley gradients. This is well illustrated in Appendix 2.

2) Surveyed terraces and straths are projected directly on to the plane of the riverbed profile wherever possible, on the assumption that the terrace profile reflects the mean path of the river during terrace formation. As this assumption is generally reasonable, the terrace fragments shown on the long profile give a fair representation of elevation and gradient, accentuated by the vertical exaggeration of scales used.

The location, tilt, and extent of these terraces are shown on Plate 7. This assumption works less well for the higher level surfaces in the centre of the gorge where strath and terrace profiles S -S3 are related to the old Waipara channel, which flowed at an angle oblique to the present channel and it is not feasible to show the trace gradients of all surfaces on the same projection.

In mapping the different terrace levels, they are first outlined throughout the entire valley, then relative heights above the present bed are used for correlation, attributing matched surfaces to the same former major bed. The numbers relate to individual terrace sequences and do not imply correlation with other rivers, the numbered sequence being from youngest to oldest. River terraces were mapped by; (1) the recognition of a sub-horizontal surface dipping downstream, and (2) identification of fluvial gravels underlying the terrace surface.

B. Results

In this study, profiles of the tilted straths related to the older channel show a general reversed tilt on the upstream side and an oversteepened gradient on the downstream side of the uplift(Sr-a.3, Sr-a.2 and Sr-a.1 profiles). Symbols are defined in Plate 8.

Where there is deformation, profiles of the terrace surface are warped, and the extent of the displacement can be determined by comparison with the longitudinal profile of the present river. For example, in the vicinity of the oblique-slip fault at the lower gorge (see Plate 8) the profiles have been significantly warped over the uplifted block, as shown on Tl-a.1 and Tl-a.2 , some to the extent that the original river slope appears to have been reversed. The backtilt of the terrace profile, correlated as Tl.10 (S68/130064), coinciding with the western limb of the actively growing Cass Anticline is quite evident. There is also a reverse fault striking subparallel to the river which duplicates the Double Corner shell beds. Bradshow & Newman (1979) described this fault as a low-angle thrust fault and infer that thrusting predated the main phase of faulting and folding, and attribute it to synsedimentary sliding. The deformation of the strath surface (Fig. 6.13) at the "Horse Shoe" (S68/134077) on the left bank of the Waipara River can be explained only by the late Pleistocene movement of this fault. The fault affected section lies on the west side of the anticline where the river and original strath gradients must be in the down dip direction and fold movements only should have accentuated the original gradient. However upstream of the fault trace the strath tilts upstream producing a divergence of old and modern profiles attaining maximum separation on the upper block of the fault.

The terrace profiles also shows a zone of upward convexity of the Tr.10 terrace surfaces (Canterbury Surface) over the projected

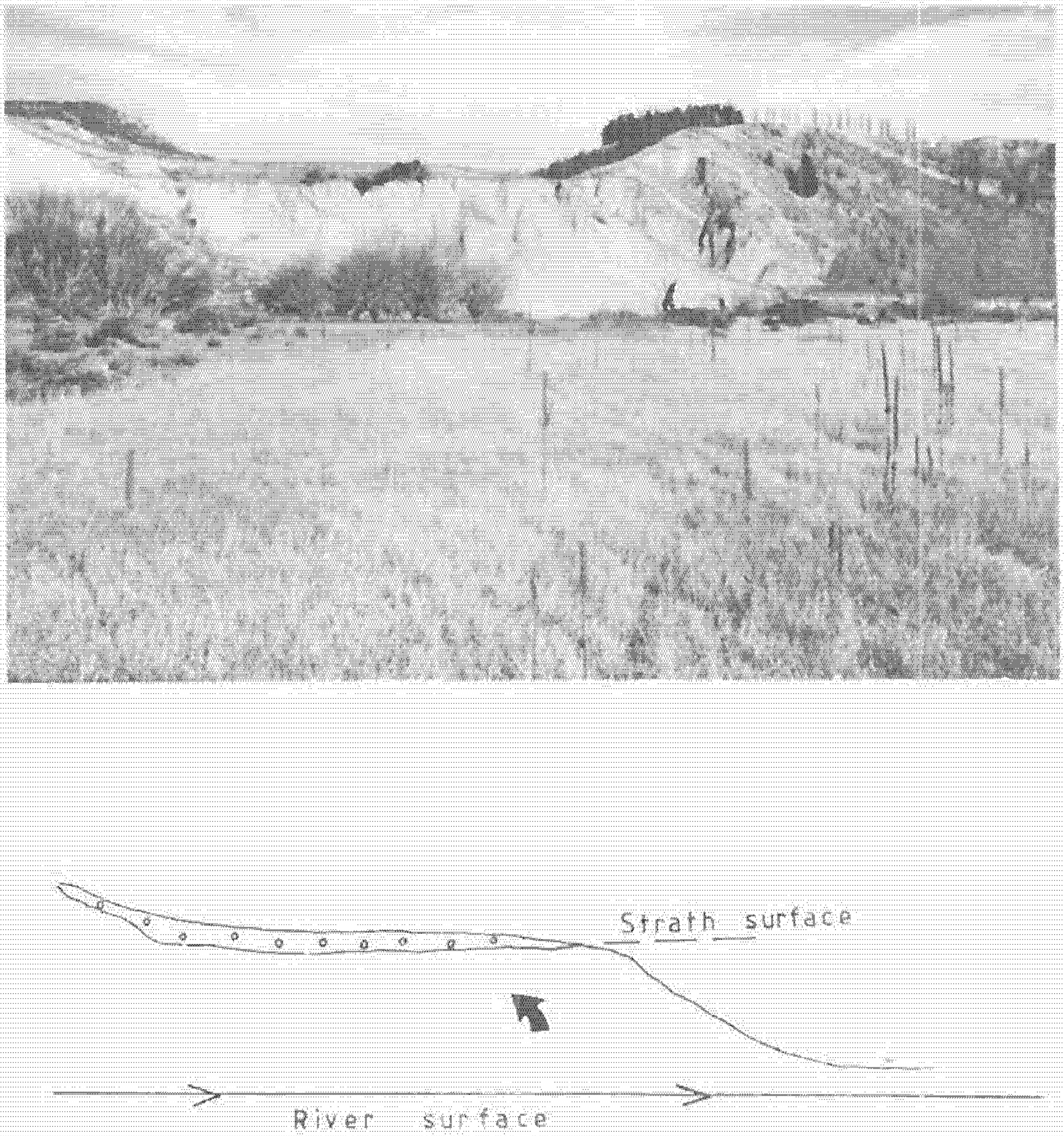


Figure 6.13

Exposure of tilted strath surface related to late Quaternary fault movement near the "Horse Shoe" (S68/134076). View looking SSW. Note the local divergence downvally between the tilted strath surface and the modern river surface.

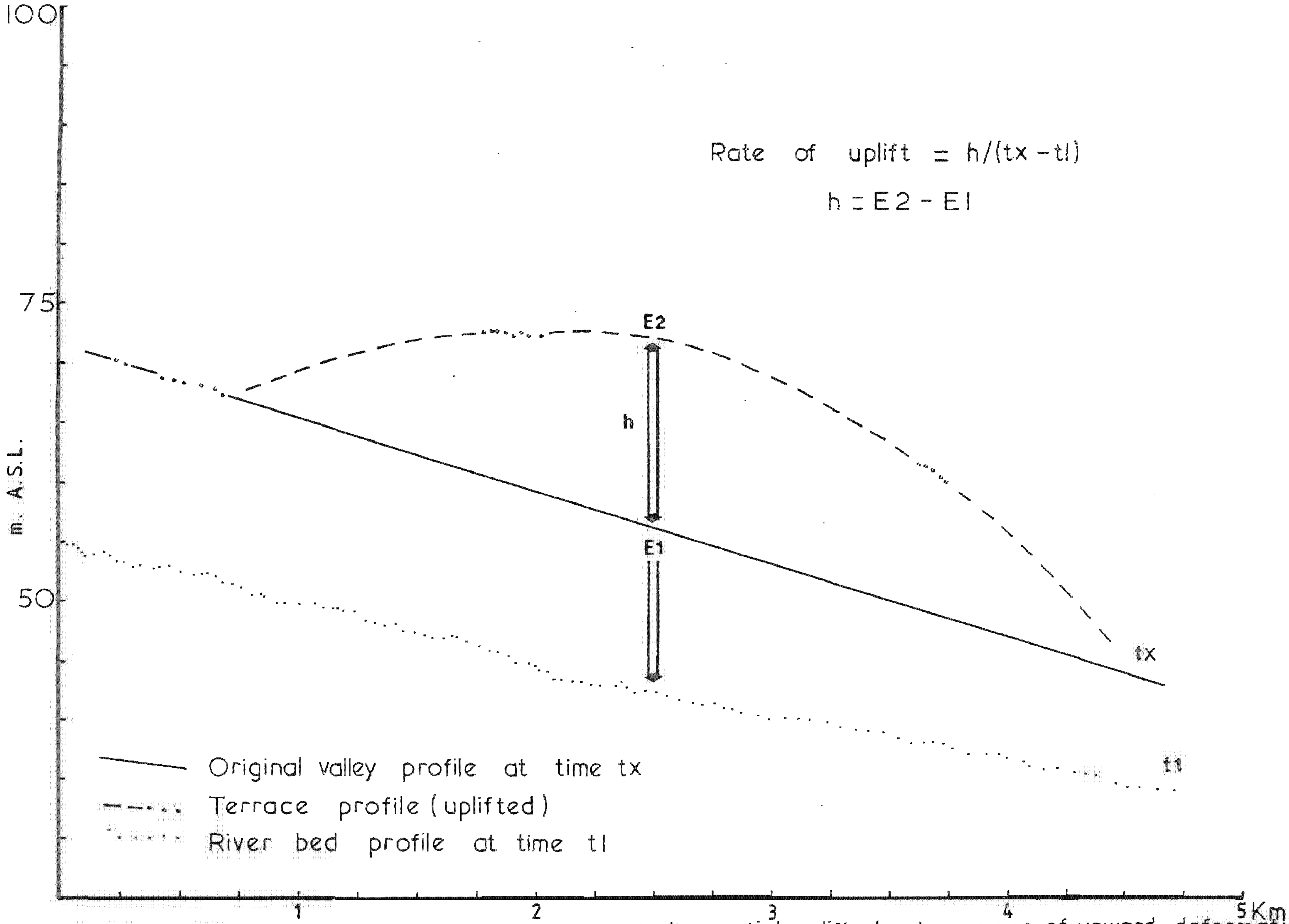


Fig.6.14 Method of estimating minimum rate of differential uplift due to a zone of upward deformation

trace of the Black Anticline. Comparing the amounts of vertical deformation of the terraces, and the river bed profile to their ages of formation provides a means of estimating contemporary rates of uplift in this area (Fig. 6.14). The amount of differential uplift can be determined as the maximum elevation difference between the actual convex profile and the river bed profile. The component-modern profile to El (h) may include uplift and/or climatically controlled base level change so that h is a measure of differential uplift only. The calculated uplift rates of differential uplift vary between 1.61 and 1.70 mm/yr for the last 10,000 years, or a little less if an older age for the Canterbury Surface is valid.

The modern Waipara River flood plain has a profile that closely matches the river bed profile. There is close parallelism between the profiles of the lower terraces (Tl.1 to Tl.5) and the present flood plain. This may be because the area is currently tectonically inactive. Precisely what initiated the downcutting is conjectural, but it might have been uplift without folding, or base level fall after the high sea level at c. 6,000 year ago, or attainment of fluvial thresholds. It is well documented (see Schumm 1977) that a series of terraces can be generated without external stimuli such as climatic or tectonic changes. Moreover, if we look at the longitudinal river bed profile of the Waipara River in detail it is broken into a series of irregular steps of alternating steep and gentle reaches known as riffles and pools respectively. The presence of the pools and riffles in both braided and meandering channels suggests that such irregularities are some kind of self-adjusting mechanism to balance energy expenditure along a stream.

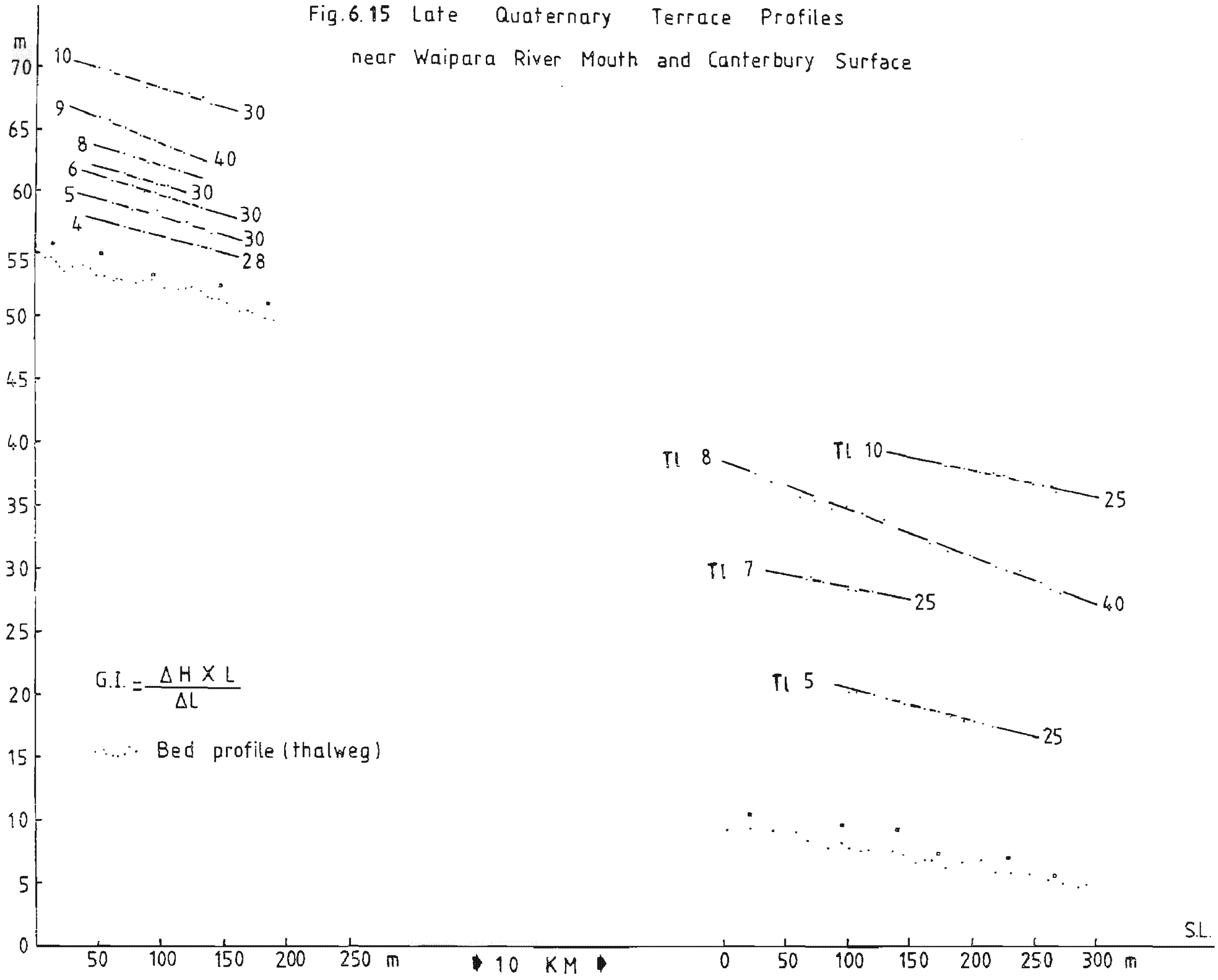
Detailed examination of the geomorphic features and strath profiles along the axis of the uplift strongly suggests that differential uplift at the south end of Cass Anticline initiated the diversion of the ancestral Waipara River from the most direct route to the sea. Such diversions appear to be strongly dependent upon the orientation and magnitude of the growth of both Cass and Kate Anticlines together with the dextral-oblique fault. It has been accepted that the successive ruptures on faults or growth events of folds associated with seismic deformation are recognized in many areas in New Zealand (Ota et al 1987, Berryman et al. 1987, Hull (in press). The anomalously high uplift rate in the lower Waipara gorge has a major control on the deviation of the Waipara River and probably could have been coseismic. There are also topographic anomalies across the Kate Anticline uplift area as recognized by the isolated remnants of upland areas near Teviotdale Homestead. As shown on the Plate 7, there is an average of 40 degrees of southward diversion from the course of the ancestral Waipara River to the present downstream orientation.

6.3.7 TERRACE PROFILES OF THE LOWER WAIPARA RIVER --A RESULT OF SEA LEVEL CHANGES OR TECTONIC ACTIVITY ?

From the detailed study of the terrace profiles (T1 to T10) shown in Fig. 6.15, terrace development must have been very strongly influenced by adjustments to both tectonic activity and sea-level fluctuations during late Quaternary time.

The gradient index (Hack 1973) of terrace profiles number 8 and 9 is much steeper than the other profiles, suggesting that the anomalously steepened gradient is a response to either tectonic activity or drop in sea level. This feature is recognisable in the same terraces near the river mouth and throughout Waipara-Omihi

Fig. 6.15 Late Quaternary Terrace Profiles near Waipara River Mouth and Canterbury Surface



valley.

Two interpretations are possible for these anomalously steep gradients. One is that (GI=40) terrace 9 was graded to a lower sea level during cooling in the early Holocene, probably at c. 8,000 years ago. This fits with a eustatic sea-level lowering at c. 8,400 years B.P., proposed by Gibb (1979). The lower terraces and modern flood plain are approximately graded to present sea level from 6.5 to 0 (Gibb 1979).

Alternatively, circumstantial evidence suggests that the anomalously steep gradient in terrace profiles may, in part, result from differential uplift. Some faults in the northwestern limb of Black Anticline and Cass Anticline were tectonically active in the late Pleistocene as the recent sediments were, in places, deformed, steepening the slope of the active stream channel. No direct evidence can be obtained to distinguish between these two interpretations if they happened simultaneously. The two terrace flights are short relative to the fold profile and they lie outside the main fold belt. In the second hypothesis, the gradient on (terrace 10), would not be affected if the base level was controlled by deformation in the intervening 10 km section.

6.3.8 TECTONIC HISTORY

At present dating of the geomorphic events in the study area is constrained only by the inferred age of Canterbury Surface and the local preservation of the late Pleistocene Tephra along the coastal section near Teviotdale. These have been estimated at approximately 10,000 and 20,000 years B.P. respectively, on the evidence previously discussed. The age of the older marine and fluvial terraces can only be estimated indirectly.

The geomorphic features reflect seven periods of tectonic and climatic base-level change related to events in the Waipara region.

1. The lower Waipara area has been particularly active during the late Quaternary resulting in the differential uplift and tilting of the last interglacial marine terraces. A period of intense tectonic activity appears to have occurred after the formation of the 80 ky terrace. This and the two older terraces, the 125 ky and 105 ky, all show evidence of tilting along the southern limb of the Kate Anticline.

2. The first aggradation period (Teviotdale Surface) accompanied or was immediately followed by the emergence and continuing strong fold movements of the south end of the central Cass Anticline.

3. The next youngest marine terrace surface formed by marine abrasion and deposition is well preserved, although the surface is now covered by secondary deposits of loess, Kawakawa ash and slope wash. The calculated uplift rate of this coastal plain suggests that uplift of the study area has been reduced during the period from the age of the 60 Ky marine terrace to that of the Kawakawa Surface.

4. The lateral migration of the ancestral Waipara river channels during and following the most active period of deformation has eroded a considerable amount of the uplifted surface. This has caused a deviation of 40 degrees southward in the mean flow direction parallel to the axis of the normal downstream migration of the meanders. The mechanism for complete realignment to a new channel in the lower reaches is the river's response to the combined differential uplift to the east and the dextral-oblique fault displacement in the

lower gorge.

The meander pattern developed as the first response to back tilting on the upstream flank of the Cass Anticline and became deeply incised with the continued acceleration of uplift rate relative to stream power; processes which are reflected in the transition from the broadly rounded older valley profile to the deeply incised meanders of the inner gorge.

5. The second major aggradation period (the Canterbury Surface) was developed during the Poulter Advance and the immediately subsequent recessional stage of the Otira Glaciation.

6. By late to post Otiran time there was evidence of reduced activity in Cass Anticline, as the present river bed has now reverted to lateral planation of the valley floor on a bed rock strath surface. Instead, deformation appears to have been taken up by longitudinal extension of the Black and Kate Anticlines propagating across the area. The deformation of the upper aggradation terrace over the adjacent Black Anticline and the anomaly in the same surface over the axial zone of Cass Anticline are indicative of late Quaternary activity.

7. The modern Waipara River flood plain grades to postglacial sea level and no evidence of deformation of surfaces probably younger than 5,000 years old has been detected in the incised meandering valley.

Surveys of shear strain rates derived from repeated re-triangulation of geodetic networks give quantitative information of the short kinematics of the shear zone. The pattern of maximum shear strain rate and the orientation of the strain ellipse, given by the direction of the azimuth of the principal axis of compression, are provided by these geodetic surveys where they are available. Within the lower Waipara area, Published estimates of shear strain rate data vary between 0.08 to 0.26 +/- 0.06 Mrad yr-1, and cover a belt that extends north-westward from Teviotdale River (Reference 22 and 23, Walcott 1984). The orientation of the principal axis of shortening varies between 100^{0} +/-7 to 102^{0} +/-20 and is , in general, consistent with the principal axis of compression postulated for the present study (see Chapter 3, section 3.1.3).

Thus, during the last 20,000 years, the modern Waipara River has created many geomorphic features showing dramatic changes of pattern and gradient in the area of the lower Waipara gorge. The uplift rates between the last 20,000 and 10,000 years, might imply that the activity of the dextral shear regime had increased over the last 20,000 years. The fact that the sites of the most active fold deformation seen to have migrated with time lends support to theoretical considerations discussed in the next case study that shear strains transmitted into Cenozoic cover as folds will be likely to produce quite rapid changes in the rates and sites at which activity occurs. During the formation of the oldest terraces, uplift was concentrated at the south end of the Cass Anticline. The younger deformation seems to indicate that the recent activity has shifted to the flank and relates to the growth of new folds in respose to accumulating shear strain in the basement.

6.4 THE CARRINGTON AND YELLOW ROSE CREEKS

6.4.1 SETTING

The "Carrington" and "Yellow Rose" Creeks, tributaries of the Waipara River in the lower Waipara area, have three well-preserved aggradation terraces of fluvial origin. The alluvial stratigraphy of the deposits that compose these terraces reflects the evolution of alluvial streams in response to climatic fluctuations and tectonic regimes of the late Quaternary. The terraces were formed by general aggradation of the valley during periods of base-level stability.

The "Carrington" and "Yellow Rose" Creeks are named informally after a local residence and have a drainage area of 2.133 km^2. They descend from an elevation of 260 and 350 m respectively on Mt Cass joining downstream to a single channel before grading to the confluence with the Omihi River at 45 m a.s.l. (Fig. 6.16).

6.4.2 PURPOSE OF THE SURVEY

The second levelling survey was made along Carrington Creek which has been incised into both the second aggradation surface and the Greenwood Formation as the result of recent warping at the south end of the Black Anticline. The reasons for selecting this particular area for the study are discussed below. From the descriptions provided, it will be seen that this area has a sufficient degree of geological and geomorphological complexity to enable an evaluation of the applicability of aerial photographs and precise levelling surveys to a study of recent surficial deformation on a small alluvial channel such as the Carrington Creek, as can be inferred from the following reasons:

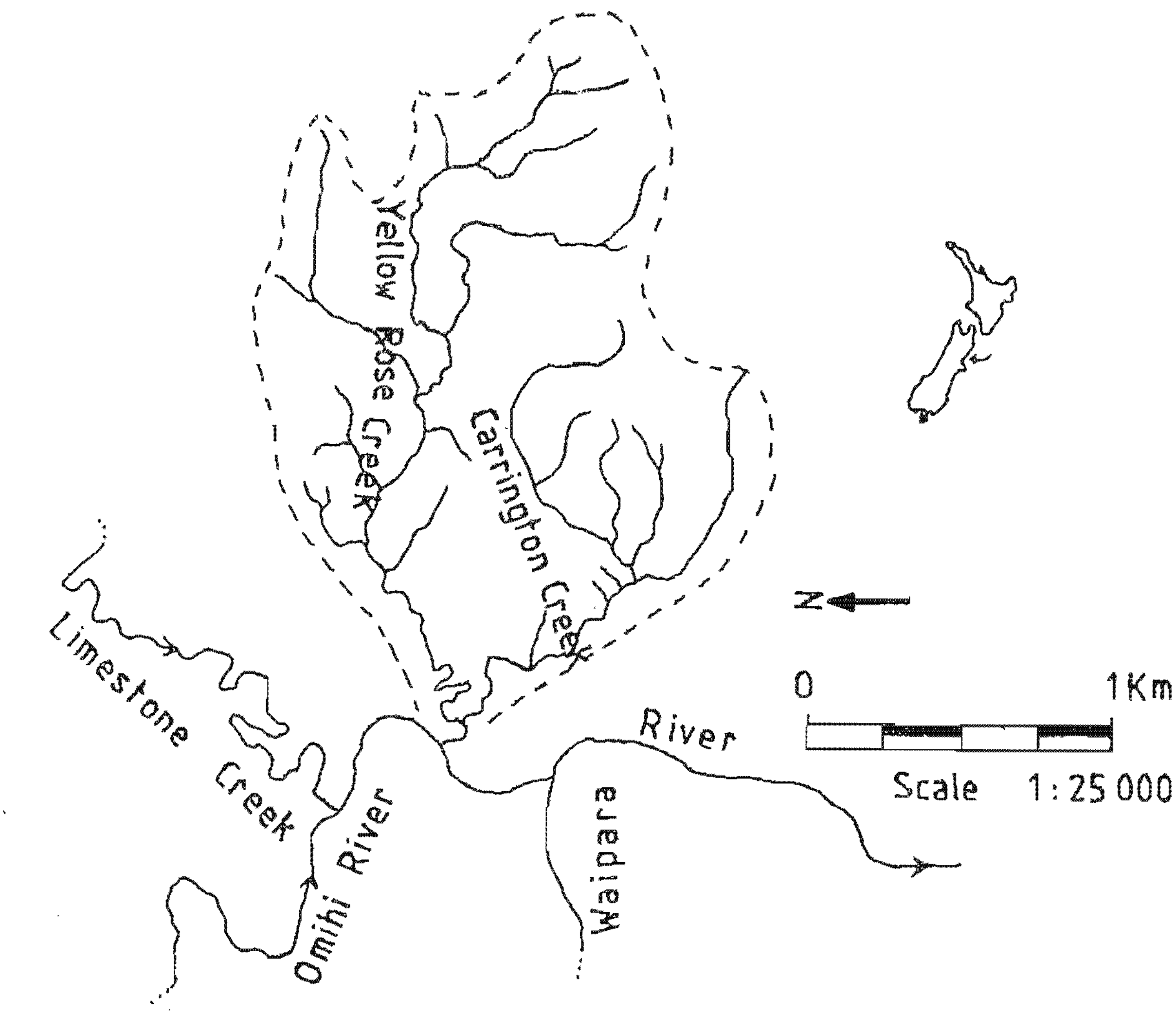


Figure 6.16

Catchment of Carrington and Yellow Rose Creeks.

1. This catchment represents a miniature or small scale example of processes which are analogous to those observed on larger scales elsewhere in the area. Therefore it makes a useful area to study in detail.

2. Despite the small size of the stream, stream power is sufficient to cross most thresholds relative to uplift and tilt, reproducing all the range of fluvial processes.

3. Data from the area adds significantly to developing a model for drainage/fold evolution etc.

Evidence for continuing effects of uplift and subsidence on channel morphology can be expressed by changes in channel pattern and sinuosity, channel aggradation or degradation, gradient, bank height, deformed strath and terrace surface, or other morphologic parameters (Schumm 1977, Schumm et al. 1982; Ouchi 1985; Yousif 1986). Integrating geologic and geomorphic data with precise levelling surveys provides a means of quantifying the rate of uplift if these data are mutually supportive. The strath surface that has been surveyed provides convincing evidence of active deformation distinguishing uplift on the anticlinal fold from subsidence in the syncline, which is also reflected by both thickness variations and deformation of the overlying fluvial deposits.

6.4.3 GEOMORPHOLOGICAL MAP

In general, the geomorphology of the area is influenced by various processes, especially gully and rill erosion, landslide, debris slides, bank collapse, slope wash, fluvial action and incision processes. These processes have, and are, still acting on soft, sedimentary rocks on which the present landscape has been developed

(Plate 9).

The geomorphic map (Plate 9) can be divided into three broad genetic groups of landform on the basis of origin, classification being based on structural, denudational and fluvial controls and processes. These morphogenetic landform units were discussed before in the context of geomorphic classification in chapter four.

6.4.4 GEOLOGICAL CROSS-SECTION

A number of geological cross-section construction methods for folds have been evolved, but the method selected should be dependent upon the competency of the rocks involved and the nature of the deformation. The "arc" method for flexure (competent) folding is adopted in this study. This method was originally described by H.G. Busk (1929). It applies only to concentric folds formed by the flexure-folding process. This method requires careful surface geological mapping of attitudes, fault-bedding angular relation ships, formation thicknesses, and small-scale structural features along the line of section as shown in Plate 10.

The cross-section is taken along a roadside cutting running obliquely up to the fold axis within the incised valley, and is 1350 m long and 5-30 m high. The cutting permits a total length of 1100 m of exposures to be surveyed and examined.

6.4.5 MINOR STRUCTURES

In addition to the Black Anticline and Cass Syncline, many minor recently active faults are present. These structures are secondary but are primarily responsible for the offset of the strath surfaces and probably play an important role in the evolution of the stream patterns, which will be discussed shortly.

The flanks of Black Anticline are complicated by the presence of a local en echelon reverse and normal faults. The en echelon nature of the crest of other folds and faults, such as those on the Cass Anticline, suggest a genetic relation to regional structures. Exposed fault planes are restricted to the river banks where they occur in both the dipping Greenwood Formation and overlying fluvial aggradation deposits. The main movement or movements appear to have taken place after the deposition of the fluvial deposits forming the second aggradation surface, but before the final stage of downcutting. Movement clearly continued until a relatively recent date but prior to deposition of at least the uppermost part of the loess capping; the significance of which will be discussed shortly.

6.4.6 LOCATION AND TIMING OF DEFORMATION

This section describes the deformation and thickness variations of the fluvial aggradation deposits and demonstrates how these data are used to determine the location and timing of deformation.

All measurements were made principally using a Wild DI3 Distomat and T1A Theodolite (Appendix 2), delineating points where both the strath and terrace surfaces are exposed or well controlled. Areas where bank collapse and landslide have modified the strath or terrace surfaces were avoided during the survey. River, strath and terrace profiles are shown as three different dashed lines on (Plate 10) accompanied by a geological cross-section.

The geomorphic terrace surfaces can be divided into three groups based on elevation above stream level. The stratigraphy and the chronology of the deposits that form the terraces will now be described in detail.

A. Upper Aggradation Surface

The oldest fluvial aggradation surface is present along the trough of the Cass Syncline. Evidence of fluvial channel deposits can be seen at the downstream end of the Carrington Creek valley (Fig. 6.18) where the synclinal structure closes round a northeasterly plunging axis. This abandoned channel indicated that the ancestral course direction drained straight into Waipara River (Fig. 6.17). This upper terrace is underlain by an alluvial fill unit which, where exposed, has the same local source materials as has the middle terrace alluvial fill (Fig. 6.19a). Current direction shows that the ancestral channel was flowing southwest near the air gap before the present northwestward swing in the realignment of the valley occurred.

This surface probably aligns with the Teviotdale Aggradation Surface near by. Their age is between 70-75 ky, based upon terrace elevation and the late Quaternary chronology previously disscussed before in chapter five (section 5.7).

B. Middle Aggradation Surface

The second aggradation surface is exposed extensively along the Carrington Creek valley (Fig. 6.19a), and because these fluvial deposits are well-exposed elsewhere (for example Yellow Rose and Limestone Creeks) and are similar in appearance and source materials, the middle aggradation surface exemplifies the process of cut-and-fill postulated for the late Quaternary in this area (Plate 9).

The lateral extent and facies changes are clearly visible. The fluvial deposits consist of at least three fining upward cycles

Figure 6.17

Landforms along the ancestral Carrington Creek valley. View looking SW. Note the valley-floor features, air gap, and the ancestral Waipara terraces (arrow) in the distance.

Figure 6.18

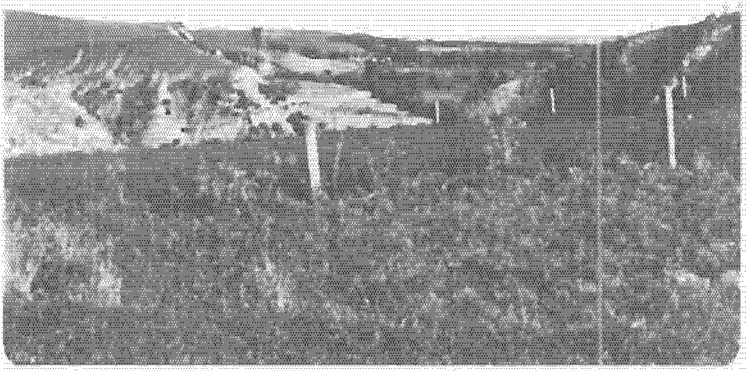
Exposure of ancestral strath surface (arrow) of the upper aggradation deposits. (Grid ref. S68/142110).

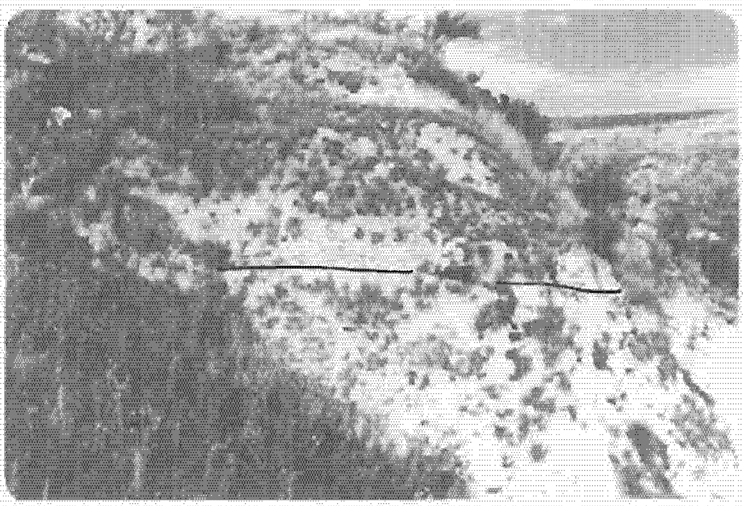
Figure 6.19a

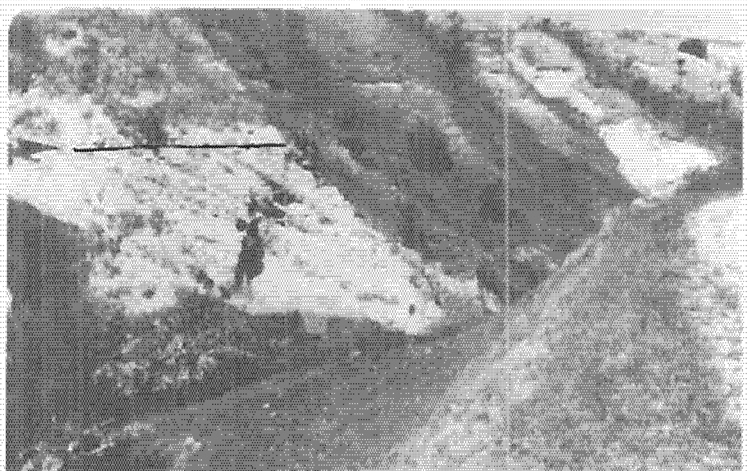
Carrington Creek bank exposure showing the middle aggradation gravels resting unconformably (arrow) on moderately tilted Greenwood Formation.

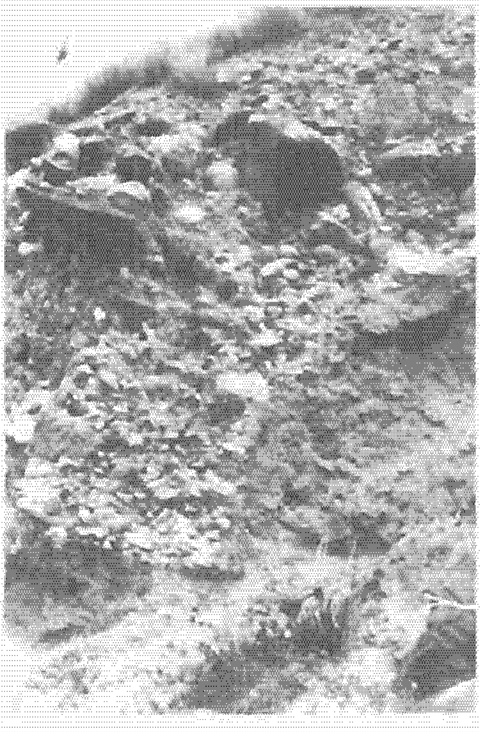
Figure 6.19b

Close up view of the middle aggradation deposits. Note poor sorting and high percentage of matrix.









covered by younger soil and loess deposits. This may be as the result of the river change in response to changing climatic and tectonic conditions.

Variations in the total thickness of the fluvial deposits provide valuable information for determining the location and timing of deformation. The fluvial deposits are anomalously thick (about 20 m) along the synclinal trough (Plate 10). Thinning occurs in two zones; along the hinge zone of Black Anticline, where it is only 9 m thick, and on the northeastern limb of Cass syncline where exposures do not permit measurement of the actual thickness.

The basal cycle above the erosional surface consists of creamy-brown, well-cemented, poorly sorted, angular to subangular gravels of Tertiary limestone clasts up to 15 cm in diameter but fining upwards into a sequence of sandstone, siltstone and mudstone (Fig. 6.19b).

Another basin on the northeastern side of the Black Anticline (Yellow Rose Creek valley) was also developing at the same time; it is similar to that of the Carrington Creek deposits based upon the lateral extent, source material and thickness of fluvial deposits (Fig. 6.27). Thickness variations in the fluvial deposits of the Yellow Rose Creek parallel that of Carrington Creek where both thin across the fold hinge.

The western margin of the middle aggradation surface occurs at the junction between Limestone Creek and Omihi River and merges with the Canterbury Surface at the Waipara-Omihi Valley (Plate 7).

A date, 895 +/- 54 yr B.P. (N34/f75) was obtained for a tree-trunk in a deposit of fluvial gravel (Appendix 3). The section shows that a sheet of alluvial gravel had slid down the Carrington

Creek valley side and the buried sample is exposed along the toe of the landslide. The date is consistent with the view that the landslide has happened after the incision takes place. There is evidence for widespread active landslides along the valley sides as shown on Plate 9. After the author obtained the date of this landslide, a second failure occurred. Heavy rainfall is likely to have contributed to the second failure.

Molloy (1977, p.161) published a series of dates on charcoal from the Cass Basin and surrounding areas, ranging from 392 +/- 37 to 789 +/- 53 yr B.P., which, he considered, probably all dated from only one fire episode. It is possible, but not absolutely certain that the date for the landslide indicated above denotes major vegetation changes which will have contributed to the stability of the slopes in the Waipara district.

C. Lower Aggradation Surface

This third aggradation surface is the youngest deposit and is present in the incised Yellow Rose Creek valley only, where it can be recognized both upstream and downstream from the Black Anticline (Plate 9). The only matched terrace is present on the upstream side, where it provides exposures from which interpretations of the extent of this terrace along the incised valley can be made (see Fig. 6.27).

A date, 1150 +/- 55 yr B.P. (N34/f76) was obtained from a piece of wood in a deposit of blue-grey, fine-grained sand and silt accumulated in the consequent drainage valley building the lower aggradation surface (Appendix 3). The sequence consists of fluvial gravel derived from Tertiary rock (mainly limestone) separated by sands and silts (Fig. 6.20). In general, these sediments tend to

Figure 6.20

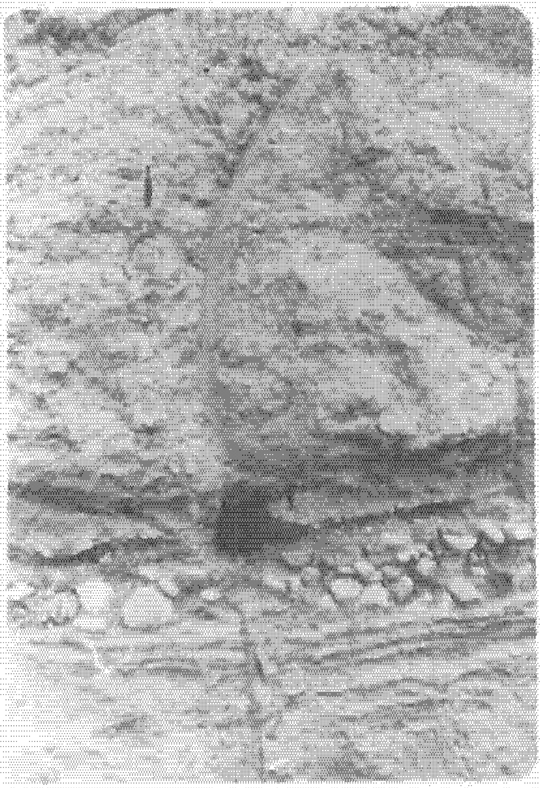
Close up view of the lower aggradation deposits along the Yellow Rose Creek. Note the fining upward cycle. Wood material was buried by fluvial deposits and dated at 1150± 55 B.P.

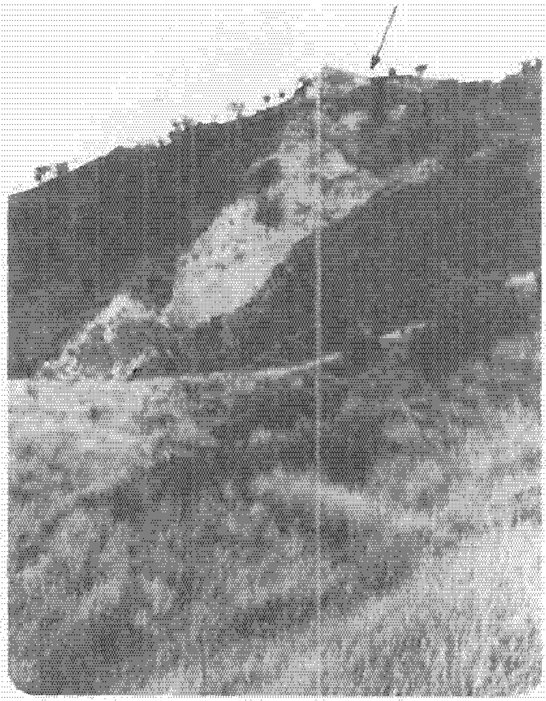
Figure 6.21

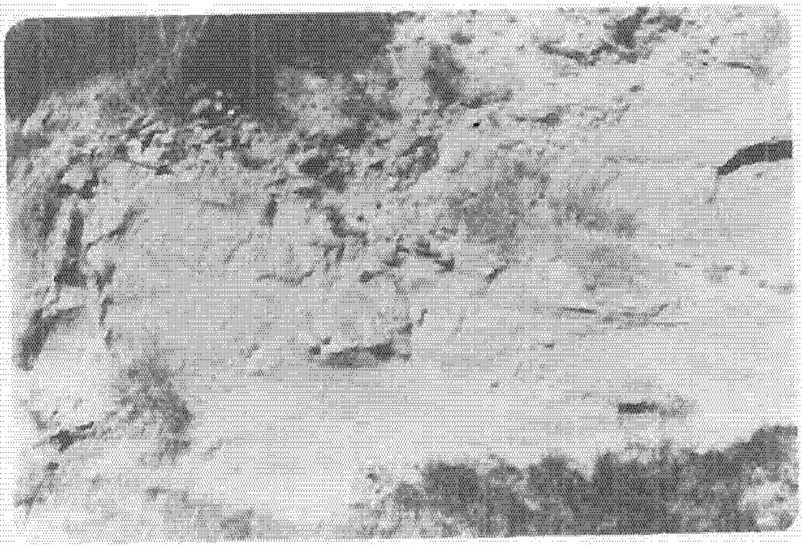
Exposure along the lower Yellow Rose Creek, looking north. Note the off-set of upper aggradation surface by reverse fault (arrow) and the cut-off channel in foreground.

Figure 6.22

Exposure of recently tilted strath surface (arrow) of the lower aggradation surface along the Yellow Rose Creek.







increase in grain-size downwards, indicating a gradational sequence developed as a result of valley filling (Fig. 6.20). The deposits are exposed along a low cliff formed by recent stream incision and show that at least partial filling (up to 4.5 m or more of sediment) had occurred before the most recent down-cutting by the Yellow Rose Creek. The date may be taken as an approximate estimate for the end of the valley filling and pre-dating the recent incision.

6.4.7 INTERPRETATION OF ALLUVIAL STRATIGRAPHY

Therefore the alluvial stratigraphy of the Carrington and Yellow Rose Creeks can be related to deposition during both periods of valley aggradation and at times of channel incision. The upper, middle and lower alluvial fill units, consisting of thick accumulations of thin-bedded, fining upward alluvial sequences, represent long-term aggradation of the creeks. The timing of periods of aggradation can be related to radiocarbon dates that bracket the alluvial fills.

At present, dating of the geomorphic events in the study area is constrained only by the inferred age of the Canterbury Surface and by the local preservation of the lower aggradation surface in the Yellow Rose Creek. Events intervening between the two periods of aggradation leading to the construction of the middle and lower aggradation surfaces are therefore bracketed between approximately ten and one thousand years respectively.

Limiting ages for the time of incision can be interpolated from the youngest age of an alluvial fill unit and the oldest age for the succeeding alluvial unit. For example, the incision of the second alluvial fill must have occurred after 10,000 yr B.P., the youngest date for the surface of the second alluvial fill, and before 1000 yr

B.P., the reliable date for the base of the inset lower alluvial fill. The timing of incision of the lower alluvial fill is more tightly constrained. Incision occurred after 1000 yr B.P., based on radiocarbon date.

The causes for the aggradation and degradation in the Carrington and Yellow Rose Creeks can be evaluated by comparing its alluvial chronology with that of other streams in the lower Waipara area. Like the uniformity observed in other studies of alluvial stratigraphy on this area, significant similarities exist in the alluvial chronologies among the eastern tributaries of the Waipara River reflecting regional events in the Waipara region.

The Carrington Creek is relatively straight where it flows along the axial region of the actively forming syncline that clearly controlled the early alignment of the creek. Evidence of uplift and steeping of the northeasterly plunge along the downstream end of the syncline exists prior to the middle aggradation surface, as indicated by the diversion of the ancestral course direction, away from the upper aggradation surface and air gap, preserved in the hinge closure (Plate 9).

Uplift was primarily concentrated along the end of the synclinal trough but the south end of the Black Anticline was also growing by at least the time of the late upper aggradation surface or early in the downcutting time for the middle aggradation surface, as shown by the discordant truncation of the strath surface across the bed rock (i.e. angular unconformity, Fig. 6.19a). The anticline clearly provided the lesser barrier to the stream than the syncline hinge at the time of diversion.

Several lines of evidence indicate that folding and faulting continued well after the deposition of the middle aggradation surface. The presence of the folded and faulted strath surface indicates that deformation continued into at least 1000 yr B.P. and is probably still active now.

Thus, the site of the most active deformation has concentrated at the south end of the Black Anticline area, and recent deformation appears to have been taken up by longitudinal extension of the Black Anticline propagating across the area in response to accumulating shear strain in the basement. The shear zone mechanisms indicate that principal compression is oriented northwest-southeast and is compatible with continued growth.

6.4.8 GEOMORPHIC EVIDENCE OF DEFORMATION

Both deformation and thickness variations of the fluvial deposits produced by fold growth associated with periodic degradation and aggradation, recorded the tectonic history of the Black Anticline. Most fluvial deposits covered the axis of uplift and subsidence, thus the strath surface at several instants of geologic time can be determined with a high degree of confidence. This record of structural relief at specific instants of time allows the rate of uplift to be calculated.

While it might be argued that thickness variation in the aggradation fill might simply reflect pre-existing topgraphy, the anomalies in the strath surface cannot be attributed to this cause (Plate 10). Thus explaining the anomalous convexities, concavities and offsets of the strath surface, which clearly relate to underlying deformation structures in the Carrington Creek valley, by late Quaternary deformation is an assumption it would be difficult to counter. The corollary, that the overlying thickness variations and the less pronounced, but still evident, anomalies in the terrace surface are also a consequence of ongoing deformation, therefore, is also a reasonable assumption. It is evident too, that the strath surface provides the best available datum surface to use for reference in determining uplift rate, in preference to the terrace surface.

Both the anticline and syncline existed prior to diversion with some synclinal closure since the upper surface is unconformably channelled into the hinge.

The geological cross-section shows that the surface expression of the Black Anticline after strath cutting was that of a gentle, low-amplitude fold, but at depth, the angular discordance of bedding against this unconformity surface shows that a substantial percentage of total shortening had already occurred, although the subsequent compression could still produce significant uplift. The fold axis runs through the centers of the convexity of the strath profile and the upper and lower limits of this convexity on the profile define the actual boundaries of recent Black Anticline uplift. Moreover, this cross-section (Plate 10) is consistent with observed faulting and it also suggests that more structural relief developed by at least late second aggradation time.

Most minor structures (i.e. faults) apparently post-date the erosional surface of the middle aggradation surface and may mark a transition from folding to faulting as the limit of fold accommodated shortening is reached. The lack of variation in aggradation thickness across these structures indicates development late in, or after, the second aggradation surface deposition. The greatest deformation occurs in the hinge zone of the Black Anticline, but, about 125 m west, folded strath surface has been offset by four en echelon reverse faults with throws ranging from 0.8 to 1.70 m (Fig. 6.23). The fault plane has a dip between 42° to 52° w, striking 010° to 020° respectively and with the reverse faults dying out away from the vertical axial plane (Plate 10). On the west side of the Yellow Rose Creek near the junction with the Carrington Creek, a reverse fault, striking 010°, offsets overlying fluvial aggradation deposits by the order of 1.5 m, and probably influences the sinuosity and valley morphology of this creek (Fig. 6.21). To the southeast, about 500 m from the fold axis, the fault zone (Plate 10), is a complex series of both reverse and normal types occurring as sub-parallel aligned sets of step faults. Reverse faults tend to be much more dominant with throws of a higher order. The fault zone has a general dip of 55° E, striking at 030° (Fig. 6.24). Folded fluvial aggradation deposits have been thrust west with throws of up to 3.0 m based on the displacement of the strath surface. Normal faults have small throws ranging from 0.8 to 1.40 m (Fig. 6.25a). The fault planes tend to be extremely straight, sharp features and generally uncemented. Since the faults occur in a sequence of well-bedded deposits, containing a wealth of sedimentary structures, any fault development resulting from differential uplift of underlying materials should be discernible from disturbances to those sedimentary structures. In fact, the sedimentary structures, for example the crudely stratified conglomerates with imbricated pebbles, are extremely well preserved throughout, but large disturbance in the pebble imbrication is apparent in the area of faulting (Fig. 6.25a & b). I interpreted these small slip surfaces as subparallel to the bedding planes and they displace the erosional surface of the terrace probably as a

Figure 6.23

Greenwood Formation and river deposits displaced 1.70 m by a reverse fault which dips 52° W (arrows). Left bank Carrington Creek.

Figure 6.24

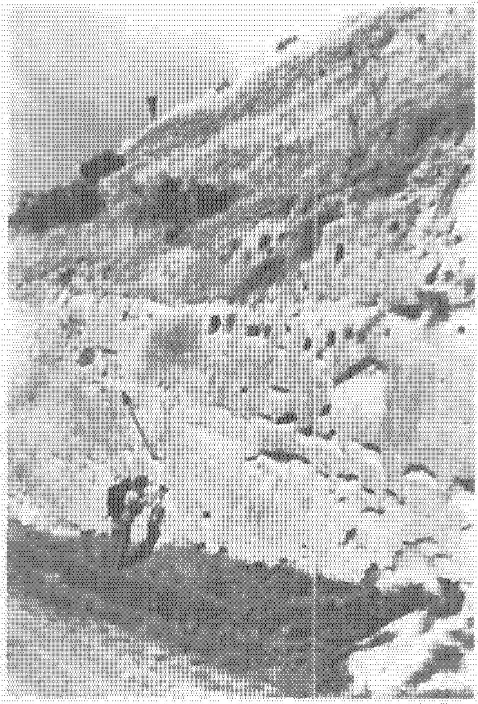
River strath surfaces displaced 3.0 m by reverse fault which dips 55° E (arrow). Right bank Carrington Creek.

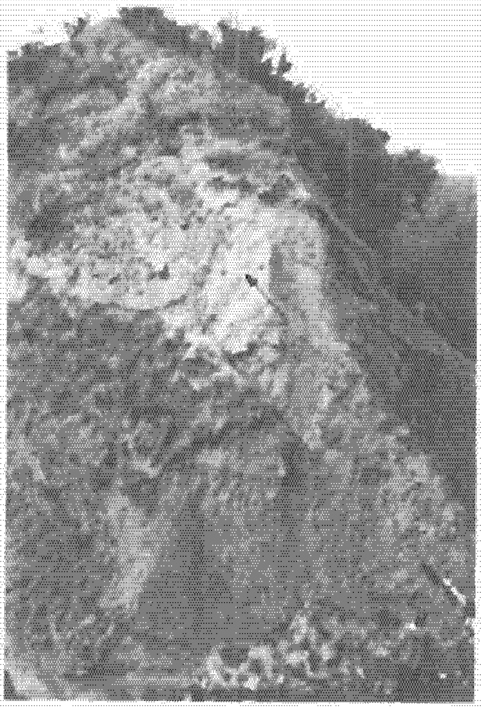
Figure 6.25a

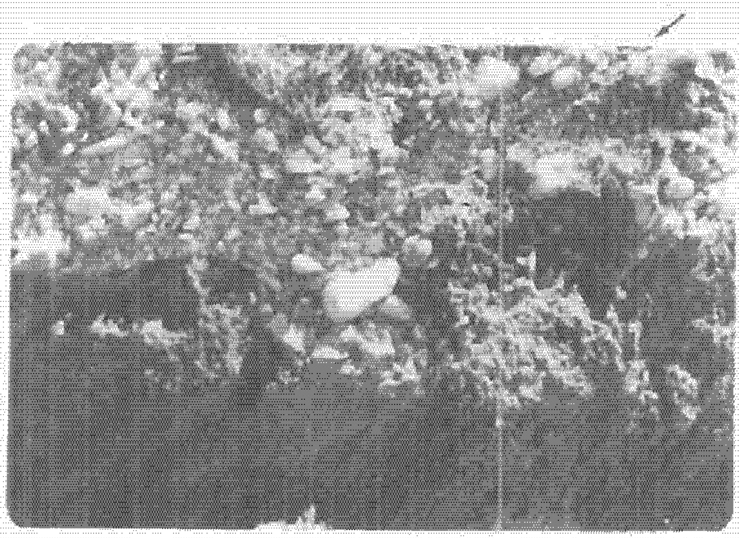
A section through a river terrace showing the strath surface cut into bedrock. Note the irregularity of the strath surface and the disturbance in the pebble imbrication caused by faulting (arrow). For more details see Figure below.

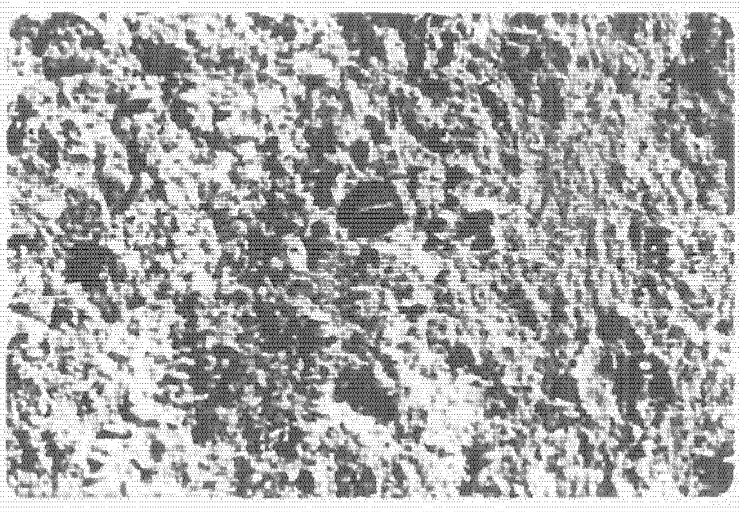
Figure 6.25b

Close up view of the conglomeratic beds showing the disturbance of pebble imbrication by faulting.









result of flexural - slip folding. The opposite sense of movement of this series of small normal faults reduces the effect of the thrusts on the terrace surface top, but a small anomaly in the terrace surface profile across this fault zone is recognized (Plate 10), despite smoothing by the slope deposits and loess cover.

These faults all lie within the Greenwood Formation and are traceable into the overlying fluvial aggradation deposits while the capping loess is less disturbed and thus masks the fault scarp so that no trace is visible on the terrace surface, either on the ground or in aerial photographs.

Because fold growth during the time of fluvial aggradation is measured by both thickness variations and deformation of fluvial deposits across the axis of uplift, each point on the strath surface of Plate 10 is assumed to represent an initial downstream grade before the warping took place. This part of the strath surface therefore reflects the fold geometry and thus provides the best available measure of the rate at which the bed rocks were being deformed.

A key point in interpreting (Plate 10) is the importance of separating gradient relief from total structural relief. During degradation of the bedrock, the gradient relief is the relief measured at the strath surface at any time, but total structural relief is an accumulation of structurally produced topographic relief throughout the growth history of the uplift after the geomorphic threshold between strath plantation and active incision has been crossed. During strath surface cut, gradient relief was always small, but total structural relief increased with continued growth of the fold. Here the amount of uplift can be determined as the maximum elevation difference between the actual convex profile and the projected original profile. If the original profile is assumed to be fairly smooth then the maximum amount of uplift can be estimated. By this method, the convexity in the strath surface profile indicates that about (18.5 m) of relative vertical deformation occurred at the uplift axis since this surface was formed. Therefore, the maximum rate of differential uplift is estimated at 1.85 mm/yr over the last 10,000 years. However, the middle aggradation surface is incised up to (36.10) metres indicating an average incision rate of 3.61 mm/yr, a rate of incision higher than differential uplift rates calculated. The pulse of aggradation from the surface was presumably in response to the building of the Canterbury Surface i.e. a climatic event and the subsequent down cutting may be both a response to uplift and to change in sediment load in the Waipara River which these tributaries adjust to as a secondary response.

6.4.9 AMOUNT AND RATE OF SHORTENING

Good exposures on both sides of the incised meandering valley allow the amount of shortening to be measured accurately. Minimum total shortening from folding plus the lower reverse faults is (111 m) as measured from cross-section (Plate 10, S-S'). If the hard sandstone bed on both flanks is taken as the limit to the original fold, then shortening is calculated to be about (0.817), based upon stretch formula.

$$S = \frac{ld}{lu}$$

The stretch of a line (S-S') is the ratio of its deformed to undeformed lengths. Stretches are always positive, even for cases of shortening (Means 1976). Only a relatively small percentage of the total shortening (i.e. folding) occurred after cutting of the strath of the middle aggradation surface (i.e. after Carrington Creek had been diverted and incised across the anticline).

Shortening is related to folding, but the relationship between fold amplification and shortening is not linear. In a general horizontal compression fold model, the fold grows vertically until steepening of the limbs results in diminished vertical growth. If faulting follows folding as in both limbs of the Black Anticline, then shortening from faulting should begin after most of the vertical uplift has occurred. Harris (1982) considered that faulting is now the dominant mode of deformation occurring within the Waipara-Omihi valley and folding has not produced any noticeable effect on the Canterbury Surface. Based on these studies I propose that the anomalous convexities and other deformities of the strath surface are produced in response to a pulse of deformation in the regional dextral shear zone particularly if folding is related to intermittent fault movement at depth (for example, see Campbell and Yousif 1985).

6.4.10 EVIDENCE OF EFFECTS OF TECTONISM ON STREAM MORPHOLOGY

If the south end of the Black Anticline has been active during the late Pleistocene and Holocene times, then evidence of particular responses of the past and present streams to the uplift activity is expected. Therefore, the fluvial morphology of the area was examined to determine whether or not such evidence of stream response actually exists in the area. The existing drainage system as well as the pattern, sinuosity and chronology of the fluvial aggradation surfaces do conform with the expected responses to surface deformation.

There is considerable evidence supporting the view that both

Carrington and Yellow Rose Creeks have had erosive capacity in excess of that needed to keep pace with lowering base levels, during part or all of their history. They are thus capable of crossing one or more thresholds, as can be inferred from the following observations:

1. The characteristically broad and rounded upper level of the valley sides of the middle aggradation surface formed by lateral planation of the valley floor on a bedrock strath surface. This is in contrast to younger inner valleys that, as a result of changing relationships of stream power to uplift rate, are narrower and steep sided (Plate 9).
2. The paleochannel valley averages about 200 m in width, about twice that of the present channel. Both modern and high terraces are made up of poorly-sorted gravel up to 150 mm in diameter. Thus, evidence exists to indicate that gravel bed channels with high specific power are not incompatible with a small alluvial channel (Fig. 6.19a & 6.26b).
3. The modern flood plain carries only a bedload veneer of gravels and recent downcutting is occuring in bedrock to form a new strath surface at the present time (Fig. 6.22).
4. Stream pattern and increasing sinuosity should be very sensitive indicators of down valley slope change. This will be considered in detail in the next paragraph.

It is apparent that different types of alluvial channels will respond differently to surficial deformation; therefore, the Carrington and Yellow Rose Creeks crossing Black Anticline uplift are deeply incised and have a minimum number of meanders of small amplitude (Stereomodel 4). The largest number of meanders occur upstream of the uplift as shown on Fig. 6.29 and 6.30, but Yellow Rose

Figure 6.26a

Synclinal reach of the Yellow Rose Creek. Note the low channel sinuosity and remnants of upper and lower aggradation surfaces.

Figure 6.26b

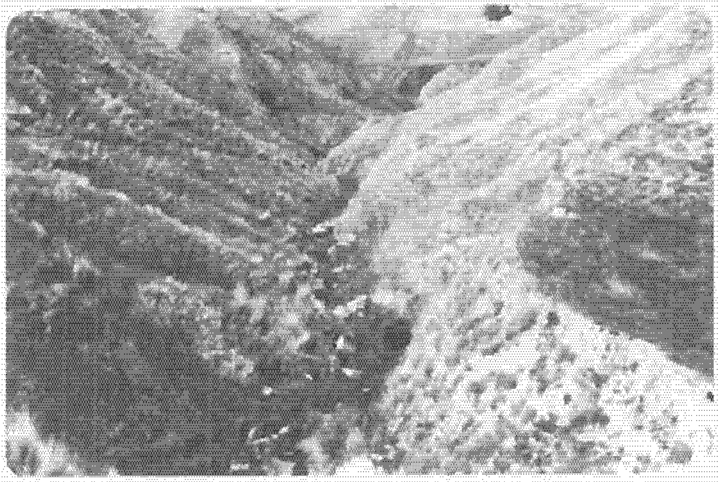
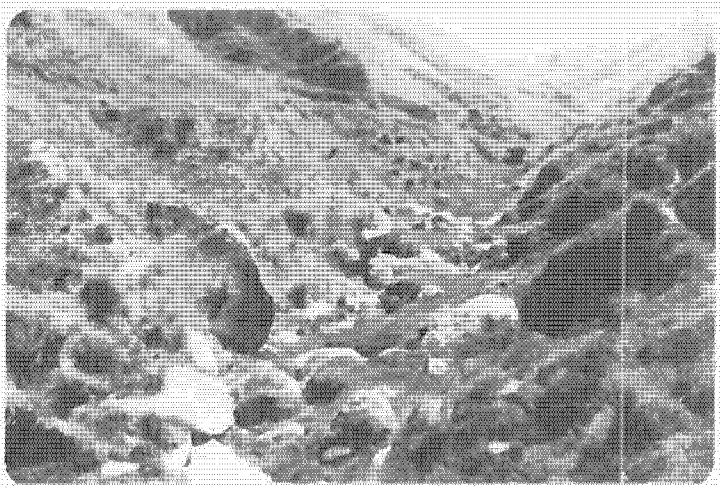
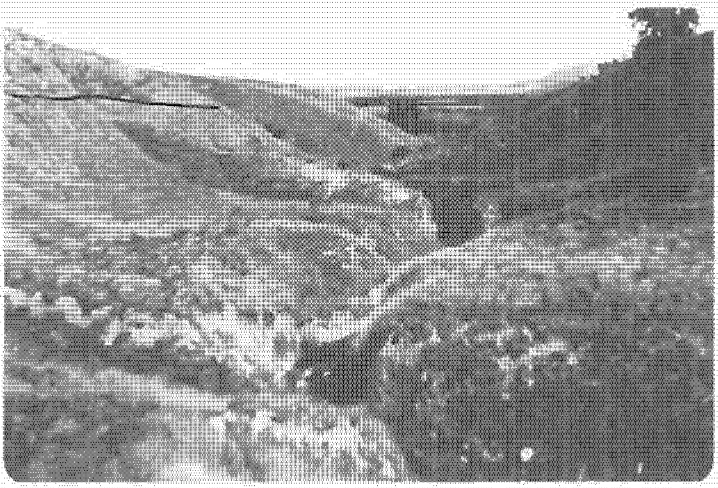
Bed load in Yellow Rose Creek. Large boulders in the river bed are common. Despite the small size of the creek it has a high bed load transport capacity.

Figure 6.27

Recent incision of the Yellow Rose Creek below the lower aggradation surface dated at 1150± 55 yr B.P. Note the upper aggradation strath (arrow) and the increased channel sinuosity compared to Figure 6.26a.

Figure 6.28

Extremely high sinuosity of the lower Yellow Rose Creek. Note the upper aggradation surface on top.


Creek shows a second high sinuosity index on the side immediately downstream. This is unlike the situation described by the author in many other examples within the study area. A clear-cut explanation for the cause of the second anomaly in channel sinuosity may not be possible because the study reach is located on a low slope section of the long profile of the Canterbury Surface where in general channel migration and meandering patterns appear to have typified the post-glacial period but, one possibility is that the meander reaches are associated with back tilting formed by the reverse faults downstream of the anticline hinge (Plate 9). Hence, the reverse faults may present a barrier with the result that upstream meanders are compressed and deformed (Fig. 6.28). The reach has a sinuosity of 3.1, and, a sequence of five unmatched erosional terraces and meander cut-offs (Fig. 6.21) were found in this recently incised valley (Plate 9). A similar pattern will result if, during incision, the river encounters resistant bedrock in a portion of its course as shown by the lower part of the Teviotdale River described in Section 6.5.6. This has also been documented by experiment (Gardner 1973,1975).

Field evidence indicates that the reconstruction of the higher level channel positions of the Carrington Creek indicates significant rotation in the general alignment of the valley axis relative to the present positions. The 90 degree southwestward swing in the realignment of the stream is quite evident. At the air gap, the old U-shaped valley is clearly seen in profile, and when projected through the upper aggradation surface matches the tilted Teviotdale Aggradation Surface, where preserved near the Stockgrove Homestead as shown on (Fig. 6.16). Such diversion appears to be strongly dependent upon the orientation and magnitude of the actively forming syncline

Figure 6.29 Relationship between river sinuosity, axis of uplift and distance of the Yellow Rose Creek.

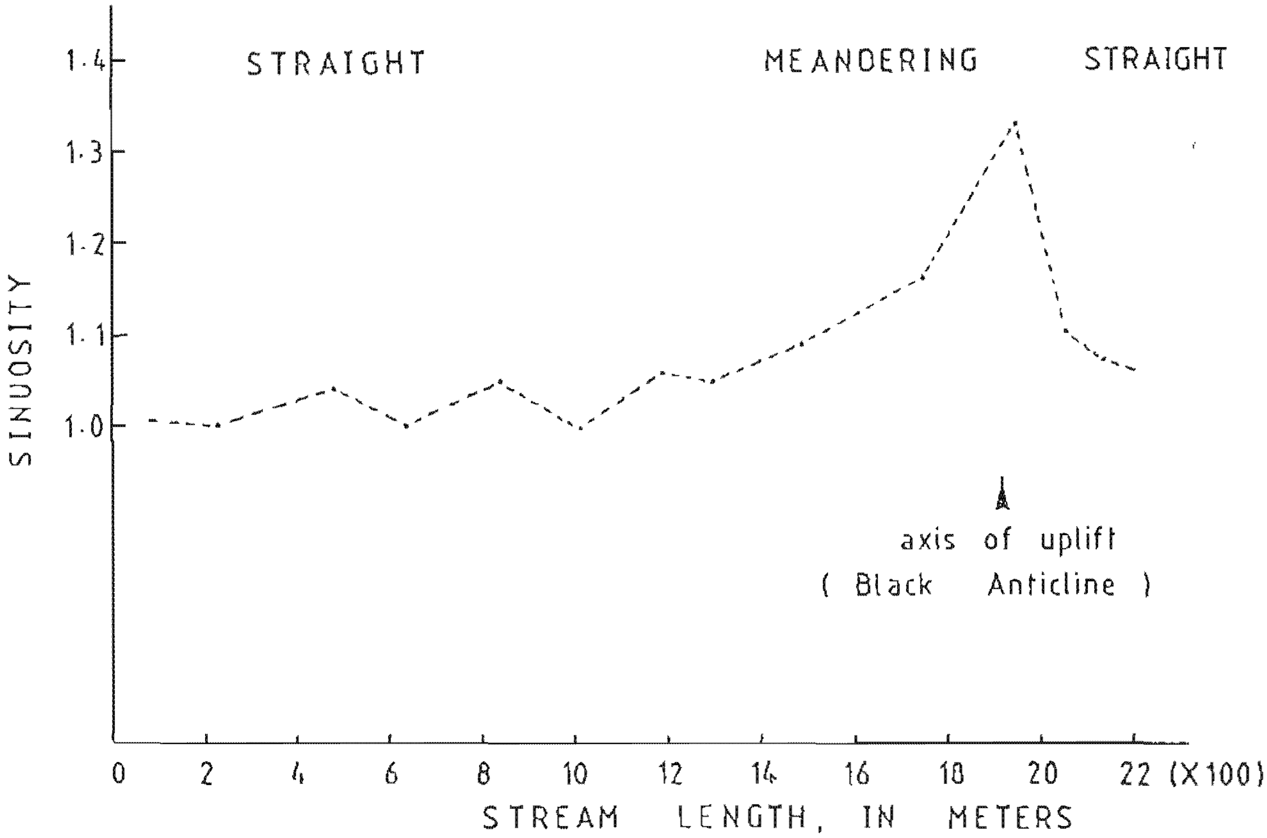


Figure 6.30 Relationship between river sinuosity, axis of uplift and distance of the Carrington Creek.

that clearly controlled the alignment of the stream, and this abrupt diversion from the old to the new channel occurs where the structure closes round the northwesterly plunging axis.

4.6.11 SUMMARY

The field relationships established between the distribution of the geomorphic elements generated by the Carrington and Yellow Rose Creeks and their interaction with the rise of the Black Anticline, suggests that a series of discrete stages can be recognised. The general model for the development of antecedent drainage on growing, plunging folds has already been introduced in chapter 6.3.4.

In this example the stream power to uplift ratio is low by comparison with the adjacent Waipara River crossing the end of the same structure. The boundaries separating each of the stages nevertheless represent crossings of geomorphic thresholds between different levels of erosive capacity. The moment the streams change for example from aggradation to downcutting, or back, may be accelerated or delayed by base level influences such as climatically induced pulses of sediment supply in the main Waipara; but the fundamental cause seems to be tectonic. Total uplift, relative uplift between fold trough and crest and tilting are the component processes. These are a response to shortening, but separating out fundamental changes in the rate of tectonic activity from the non-linear relationships of a parameter such as tilting which is inherent in the folding mechanism is difficult. Each stage in the landscape evolution must be bracketed by a component of the total shortening strain. The stages are summarised as follows:

a) Stage I: Initial slow growth of folds. Consequent drainage flows down the Cass syncline hinge directly into the Waipara River at the time of Teviotdale Surface aggradation.

b) Stage II: Accelerated closure and back tilting of the syncline accompanied by some downcutting and channelling of the stream into the Teviotdale Surface and bedrock.

c) Stage III: Diversion of the stream across the Black Anticline with ample erosive capacity to cut a broad strath-floored valley. Stray angular unconformities of bedrock against this erosion surface shows that the folding of the anticline had already advanced a long way towards locking up point when diversion of the river across it occurred. Yet the closure of the adjacent syncline must have been at a higher elevation than the crest of the anticline at that moment or the stream would not have spilled across.

d) Stage IV: Aggradation of the Canterbury Surface accompanied by ongoing folding and incipient faulting. This appears to be a climatically induced stage but once aggradation stops and downcutting begins again, the stream incises deeply, cutting through the aggradation deposits into bedrock.

e) Stage V: This downcutting represents the next phase and must represent a much more accelerated uplift than Stage III because the stream shows no evidence of any capacity for lateral planation. Is this a reflection of an acceleration of overall uplift rather than differential movement between anticline and syncline? The influence of the folding can still be observed in the entrenched meanders developed on the upstream side of the anticline in response to backtilting. The top of the aggradation surface shows signs of warping also indicating that some folding continued during this phase.

f) Stage VI: Reverse faulting on the fold limbs with modification to the streams adjusting to fault control marks the most recent phase and may indicate that the fold is close to locking.

Relative to the total finite strain indicated by the bedrock fold profile, the succession of events summarised above appear to have taken place during accumulation of a very small proportion of the total shortening strain, because even the oldest erosion surface is in strong angular unconformity on the fold limbs. They also occur close to the end of fold development, not the beginning. These comments also hold for the much larger Waipara River. If diversion of rivers across already well established structures is possible, without involving superposition, then traditional views on the nature of antecedence may be in question.

6.5 THE TEVIOTDALE RIVER AND KATE CREEK

6.5.1 SETTING

The Teviotdale River and Kate Creek both rise within, and follow, the synclinal trough of the Teviotdale Syncline at c. 275 m and 310 m respectively, and are separated by a watershed in the middle of the trough labelled "synclinal divide" on Fig. 6.31.

The Teviotdale River flows westward, and the Kate Creek eastwards along the axial region of the actively forming syncline which clearly controls the alignment of the streams.

A comparison of the behaviour of these streams at the points where they cross the axis of the Kate Anticline displays anomalies in the valley and channel characteristics which provide evidence in support of a history of antecedence across an actively forming fold (Stereomodels 18& 19). Changes in the extent of fluvial aggradation / degradation, terrace gradients and the irregularity of the long profiles possibly indicate the influence of differential tectonic movements, rather than sediment supply changes (Campbell & Yousif 1985).

Determination of the pattern and amount of absolute uplift for the Kate Anticline is a complex process involving an interrelated study of sea level changes, climatic fluctuations, marine terracing, landform morphologies and the chronology of the development of the drainage pattern.

The aim of this case study, is firstly to map the distribution and age of late Quaternary surfaces in the lower Waipara area (i.e. between the Teviotdale River and Kate Creek) with particular emphasis on the influence of structure and climatic control upon its fluvial

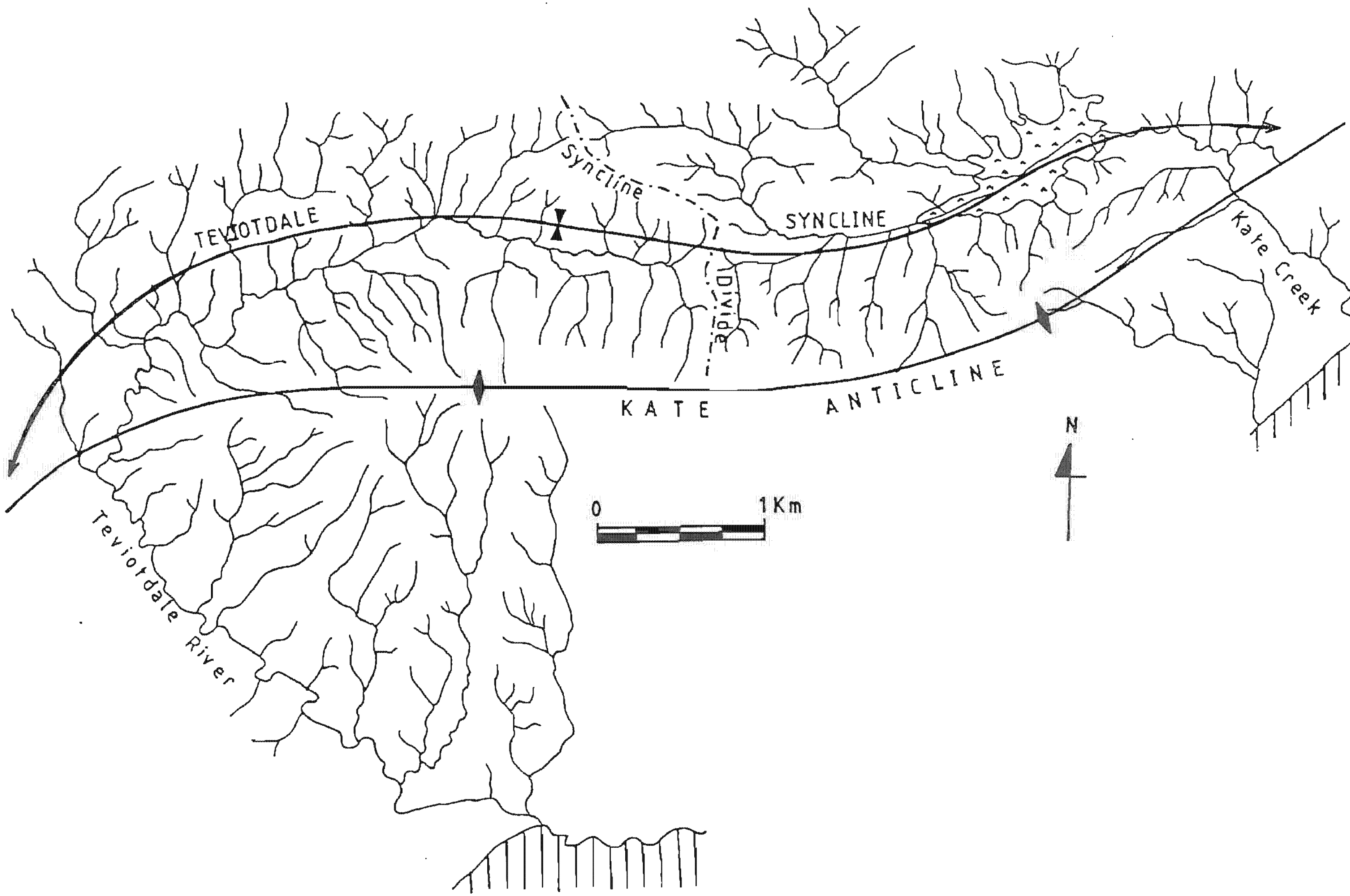


Figure 6.31 Catchment of Teviotdale River and Kate Creek with trends of major geological structures.

chronology and secondly, to use these data to examine tectonic development during the last 125,000 years.

The Teviotdale Catchment has already been introduced in Chapter 5 (section 5.5.2) as one of the case studies dealing in detail with the interpretation of stream gradient data. Here the general geomorphology was described in some detail and again considered earlier in this chapter (Section 6.3.2) in the context of the Waipara River. Discussion here will be restricted to those aspects where a useful comparison can be drawn with Kate Creek, both catchments involving development of the same underlying fold structures in common.

6.5.2 THE GEOMORPHIC MAP

In general, the geomorphology of this study area has been influenced by a number events, especially fluvial and marine processes (Plate 11). These processes have been and are still acting on the actively forming folds on which the present landscape has developed.

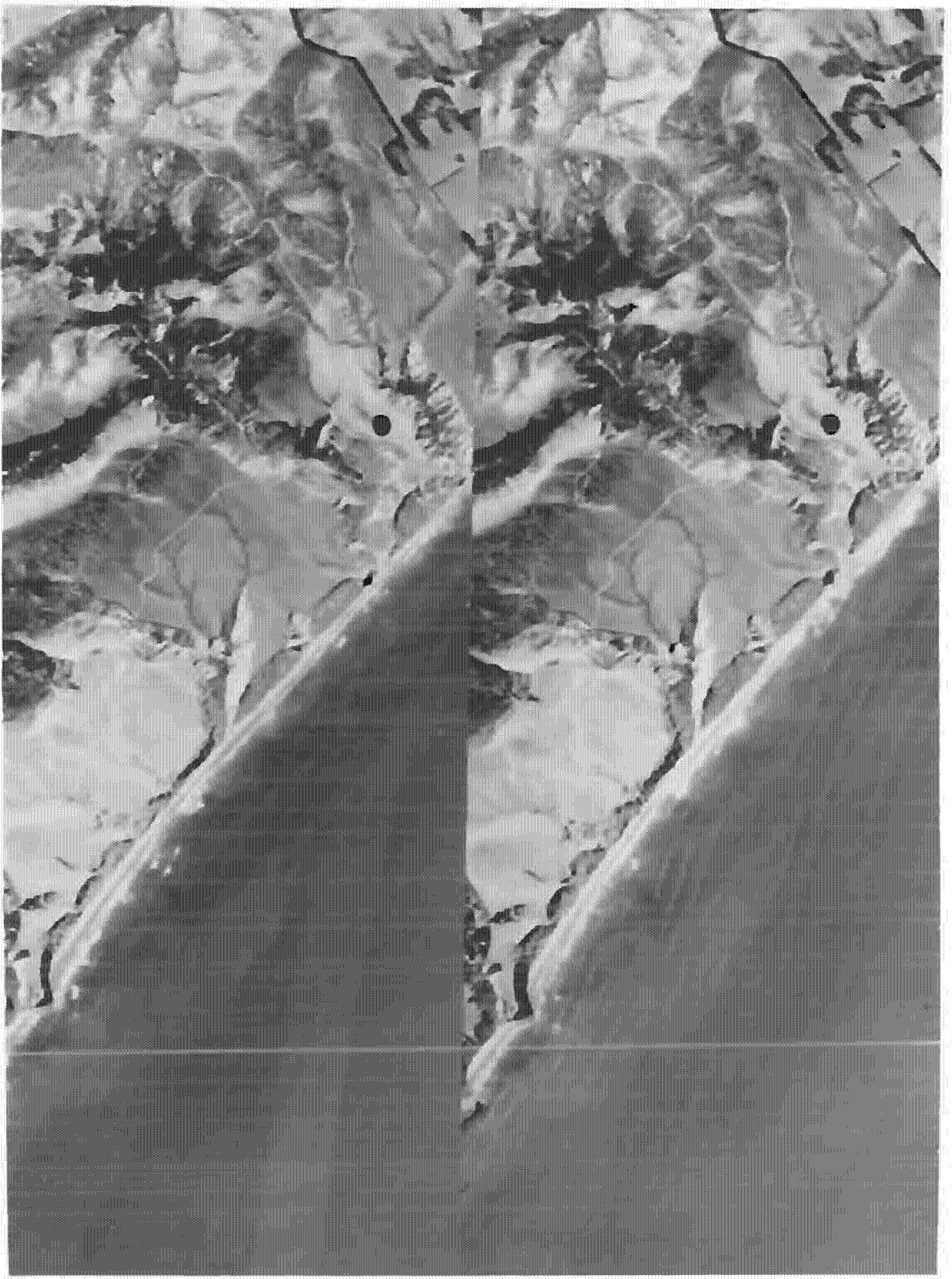
The deformation that generated the Kate Anticline is still apparent in the surface morphology as a young, unbreached anticline. Constructional and erosional processes of both fluvial and marine origin operating before, during, and after the most intense periods of upwarping, have resulted in the type of landscape shown on Plate 11 (see also Plate 2).

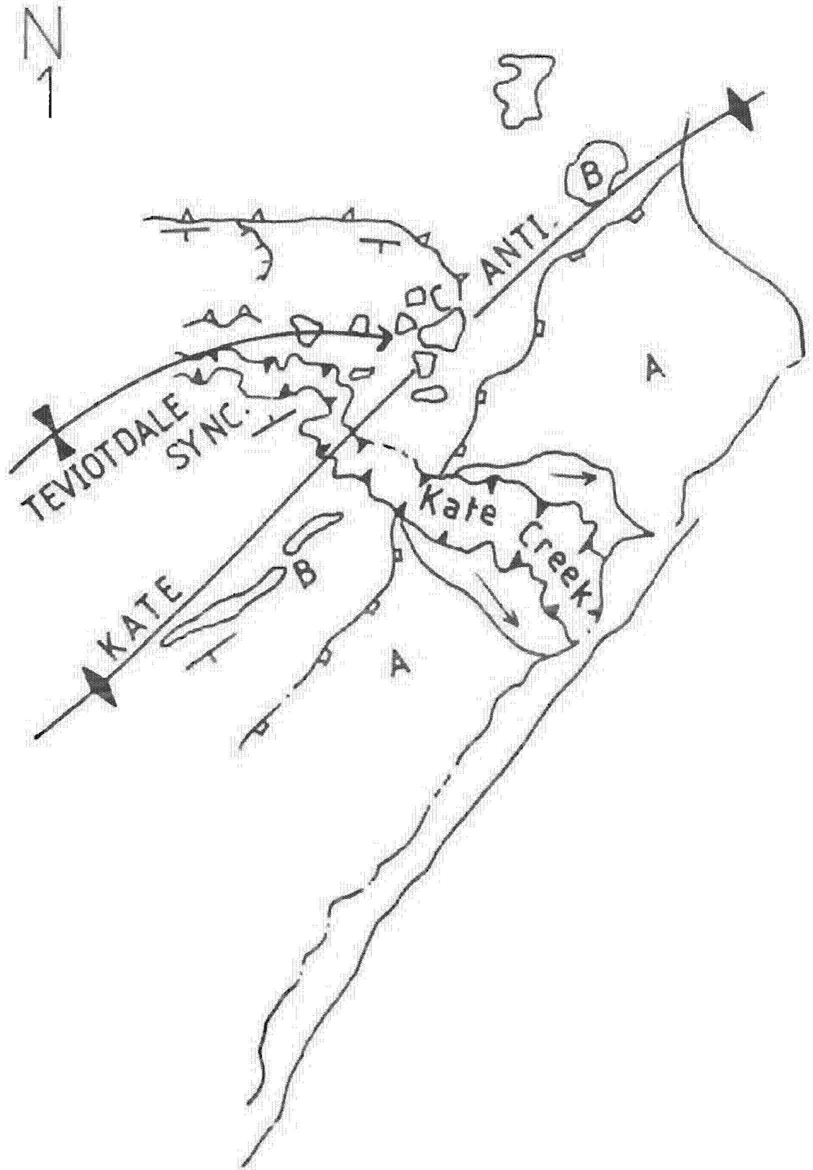
6.5.3 GEOMORPHIC FEATURES

The geomorphic terrace surfaces can be divided into eight levels based on their elevation above sea level. The stratigraphy, chronology and origin of the deposits that form the terraces in the

Stereomodel 18

Aerial view of lower Kate gorge, cut through the northern end of the Kate Anticline. Note the ancestral Kate Creek terraces near the plunging synclinal closure (area C). Flights of continuous (area A) and discontinuous (area B) marine terraces can also be distinguished.





N
TEVIOTDALE
SYNC.
ANTI.
KATE
Kate Creek
A
A
B
B

vicinity of the Kate Creek valley will now be described in detail (Plate 11).

A. The Last Interglacial Marine Terraces

Three flights of marine terraces of supposed, Last Interglacial age occur in several localities on the southern slope of the Kate Anticline (Plate 11). These former shorelines which occur beneath the back edge of each terrace mark the highest point that sea level attained for a particular eustatic sea level (see Plate 4).

The marine terraces in the Dovedale River mouth area north of the Kate Creek were classified by Carr (1970). These are, in descending order of elevation, the Leonard, Bobs Flat, and Tiromoana Formations. The stratigraphy and chronology of these terraces have been discussed previously in Chapter Two and Five respectively.

The terraces bordering the Kate Creek gorge are here referred to as erosion terraces as no marine deposits have been found on them (Fig. 6.32). The main reason for describing them as marine erosional terraces is that they occur in step continuation with lower evidently marine terraces (i.e. the Tiromoana coastal plain, M1) and at the same relative heights as the marine terraces nearby. A possible explanation for the absence of marine deposits is that there were no nearby sources of supply of such material. A possible alternative, therefore, seems to be that these could be planation surfaces formed under subaerial weathering and erosion. Located just south of the Glenafric Homestead (Plate 11) is a small flat area which is separated from the main surface, of the Tiromoana coastal plain (M1), by a steeply rising slope. Situated at the 122 m level it is a remnant of the older Bobs Flat coastal plain (M2), although there are no

Figure 6.32

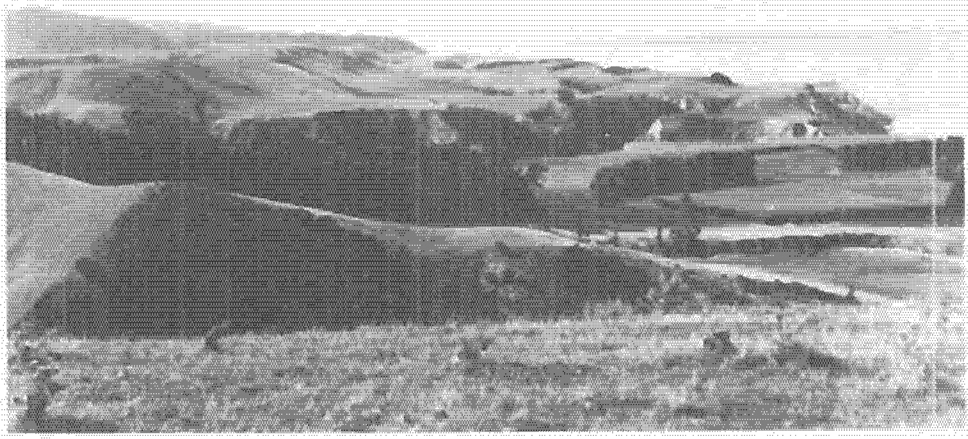
Coastal landforms north of the Kate Creek gorge. A series of last interglacial marine terraces is fringed by a last glacial terrace (Tiromoana coastal plain). At the gorge notches cut into Kate Anticline, a correlative of terrace in background.

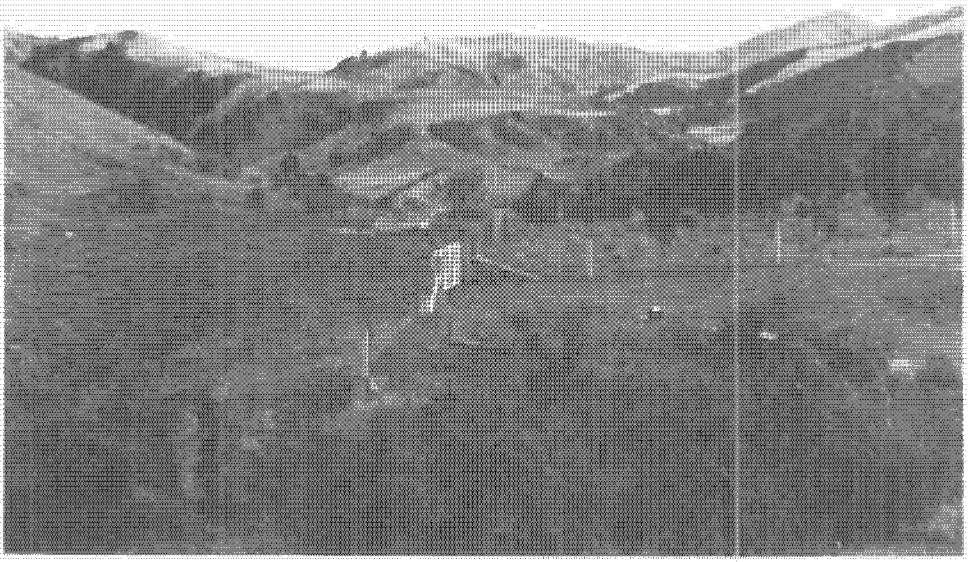
Figure 6.33

Landforms along the ancestral Kate Creek valley. View looking south-west. Note the valley-floor features, air gap, and the ancestral river terraces upstream.

Figure 6.34

Landforms along the Kate Creek gorge. View looking east. Note the incised meander valley and flights of erosional terraces.





exposures of marine materials beneath the surface to confirm this supposition.

B. The Ancestral Kate Creek Terraces

As a result of tectonic activity, coastal rivers and creeks degraded their courses and were entrenched into the raised coastal plains. Field observation indicates that ancestral Kate Creek terraces can be recognised near the plunging synclinal closure (S68/222086). Both these and the marine terraces bordering the ancestral course were formed during the same periods of high sea level during the last interglacial (Plate 11).

Flights of narrow, discontinuous erosional terraces and an air gap have been distinguished breaching the synclinal hinge zone (Fig. 6.33). It is believed that these terraces are the result of fluvial degradation which have been graded to the levels of the last interglacial marine benches formed about 125,000 to 80,000 years ago and at that time drainage was still aligned along the axial trace, before the southward swing in the realignment of the valley occurred. The present Kate Creek is deeply incised in a narrow gorge below these terraces (Fig. 6.34) where the lower course diverts abruptly across the synclinal limb and nose of the Kate Anticline.

C. The Last Glacial Marine Terrace

The Tiromoana coastal plain (M1) descends variably from the Dovedale Stream to the Waipara River, and depending on its orientation with respect to the sigmoidal axial trace of the Kate Anticline, varies from about 100 m to 38 m above sea level (see Plate 11). The coastal plain is dissected by active rill and gully erosion which has created an irregular topography. The Tiromoana coastal plain is widest at the mouth of the Dovedale River, and narrows progressively towards the mouth of Waipara River, where it is less than 250 m wide. It has been cut almost wholly into the gently inclined Tokama Siltstone and Teviotdale Gravels.

D. The Kate Fan Surface

A narrow aggradational surface bordering the Kate Creek mouth area (Fig. 6.35), falling from some 75 m to about 70 m above sea level as shown on the map, is here referred to as the Kate Fan surface (Plate 11). The Kate Fan section (Fig. 6.36) begins at the basal unconformity with an average 3 m of greywacke beach gravels or interstratified well sorted grey sand and beach gravels at the apex of the fan (S68/220078). Marine shell fragments are very common here and massive thick structureless yellow-brown, muddy, fine silt completes the section. It is highly calcareous and on exposure weathers to give a vertically fluted surface. Yellow fine sandy silt and mud dominate the seaward edge section. A river origin is suggested for the upper massive muddy fine silt where beach sediments containing shells and discoidal greywacke pebbles pass up into silts bearing no trace of marine origin. However, this narrow aggradational surface which marks the limits of the Kate Fan surface, is taken to be significantly later than the Tiromoana terminal aggradation (M1), and demonstrates a considerable period of aggradation prior to fan incision.

E. The Lower Aggradation Surface

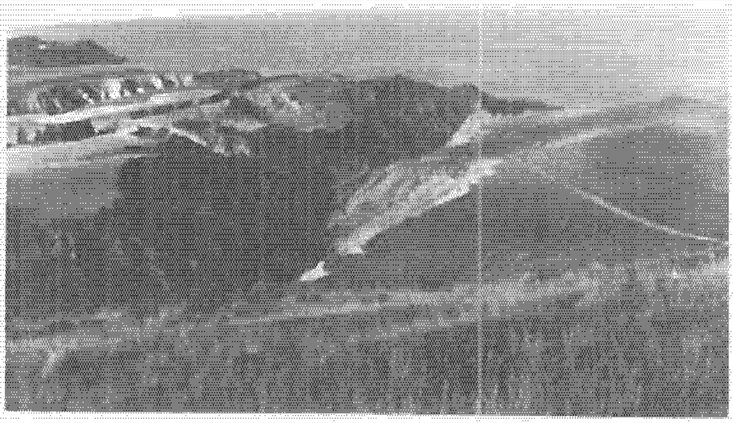
A further incised river aggradation terrace has been recognized below the Kate Fan surface at about 56 m above sea level (S68/224075).

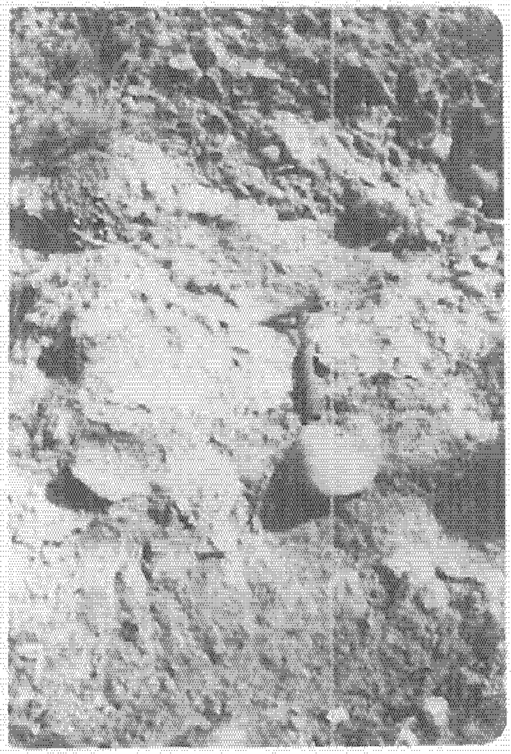
Figure 6.35
View of the lower Kate Creek.
Note the incised straight channel
downstream from the gorge and the
highest marine terrace in foreground.

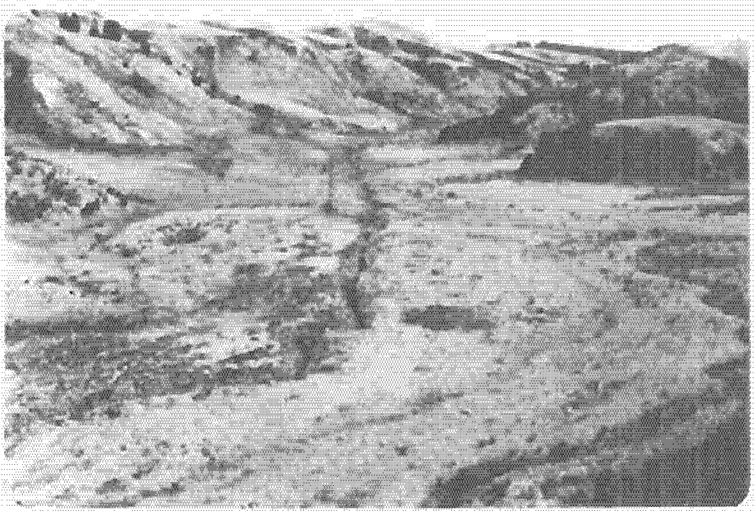
Figure 6.36
Exposure of beach deposits of the
greywake gravel facies at S68/220078,
lower Kate Creek section.

Figure 6.37
Exposure of fluvial gravels at
S68/224075, lower Kate Creek
section.

Figure 6.38
Synclinal reach of Kate Creek showing
a wide swampy aggradation surface in
foreground, dated at 775+/- 55yr B.P.







The basal part of the terrace is a 1.5 m layer of iron-stained sand and gravel (Fig. 6.37). Here the strath surface is resting unconformably upon the Tokama Siltstone. Sandy gravel exposures are largely composed of lenses of stream gravel set in medium sand. Pebbles are mostly subrounded greywacke, up to 5 cm in diameter and are found together with broken shell fragments both of which are derived from the Greenwood Formation. A fluvial origin is suggested for the iron-stained sandy gravel. It is unlikely to be a beach gravel in view of the unit's poor sorting, poor stratification, and lack of discoidal beach pebbles. It is reasonable to assume that the cut surface has a stream origin indicating a relative base-level fall caused by eustatic sea level lowering and/or tectonic uplift subsequent to the Kate Fan aggradation.

F. Holocene River Erosion Terraces

Terraces within this age group form the majority of those preserved in the Kate Creek valley. The Kawakawa Tephra has not been found in the coverbeds.

Extensive terrace surfaces are not well preserved within the Kate gorge. The greatest number of river terraces is preserved near the sea (Fig. 6.35), where four erosional terrraces are preserved from 2 to 25 m above sea level.

G. The Kate Swamp Surface

A wide aggradational swampy surface occupying the core of the Teviotdale Syncline is well preserved west of the Kate gorge (Fig. 38). Two dates, 775 +/- 55 yr B.P. and 943 +/-55 yr B.P. (N 34/f73 and N 34/f74 respectively) were obtained from two pieces of wood from a swampy fill of carbonaceous fluviatile silts and sands (Appendix 4). The sample site lies in the axial region of the Teviotdale Syncline. The samples date the end of the aggradation related to backfill and uplift. The first date may be taken as an approximate estimate for the end of the swampy filling and pre-dates the recent incision by the creek.

H. The Modern Coastal Plain

This unit occupies a very narrow coastal strip lying between the present shoreline and the postglacial marine cliff (Plate 11).

6.5.4 THE LONGITUDINAL PROFILE OF KATE CREEK

The long profile of Kate Creek from the headwaters to the sea (Fig. 6.39) is divided into three distinct sections which are separated by two knickpoints at 600 ft and 400 ft.

The first knickpoint at 600 ft, can be attributed to lithological changes, within the same formation (Greenwood Formation), a situation similar to that described previously for the upper reaches of the Teviotdale River (Chapter Five, section 5.5.2).

The long profile of the middle section, occupying the axial region of the Teviotdale Syncline, has a gradient index of 0.9 which indicates that this section is aggrading more rapidly than the upper and lower sections. Aggradation, which started in the lowermost reach of the synclinal trough, where a slope discontinuity was formed by backtilt and uplift, migrated upstream into the reach where subsidence had occurred.

Anomalously high gradient index values (GI/K = 10.8 and 9.8) in the lowest section have probably occurred in response to rapid sea

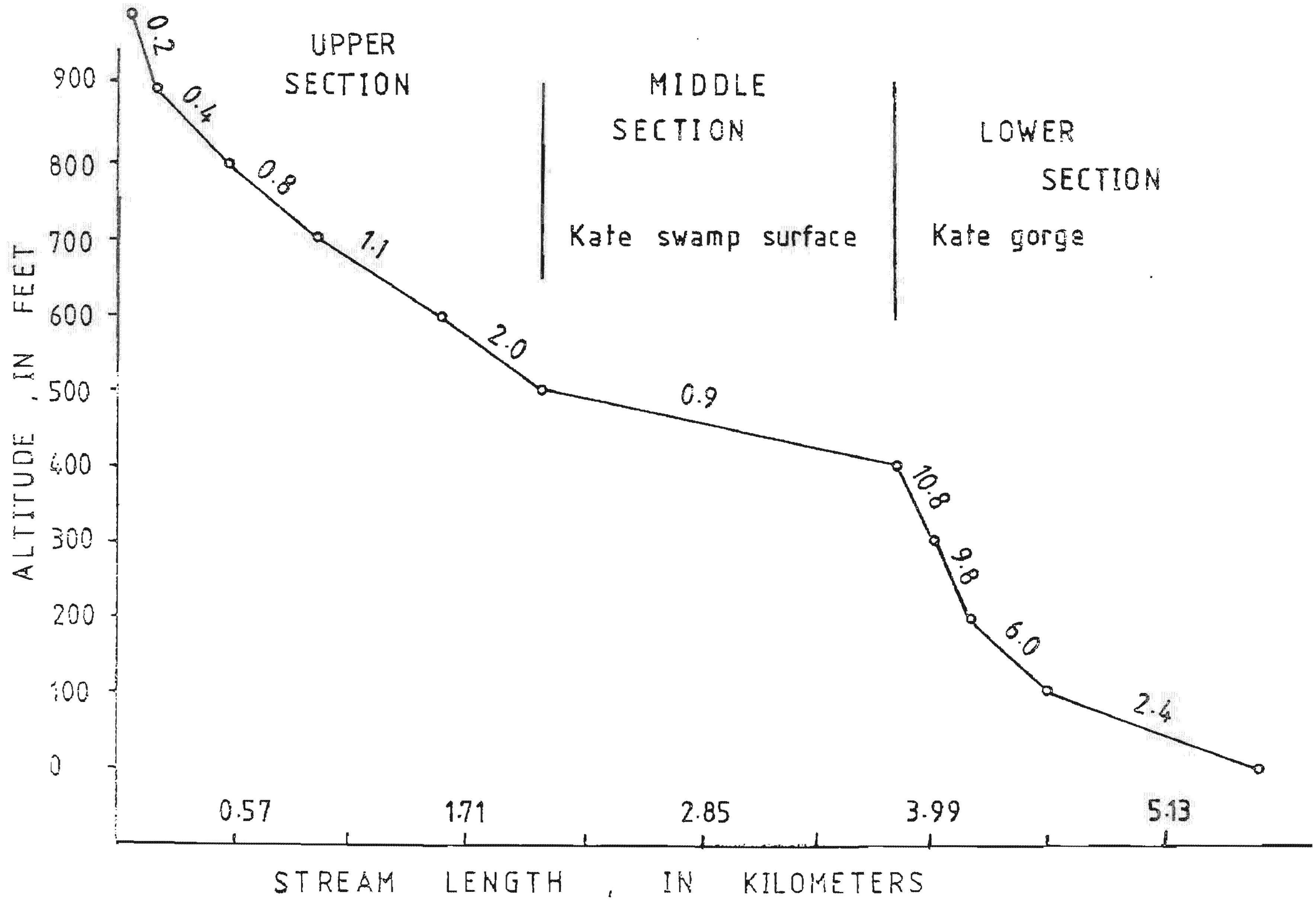


Figure 6.39 Longitudinal profile of the Kate Creek.
Numbers on river represent gradient indes (GI/K).

level change and differential uplift, both of which factors will be discussed in the next section.

Active cutting of bedrock of Upper Tertiary age across the axis of the Kate Anticline by the present creek provides further evidence of active deformation. A number of low level terraces (Plate 11) which are graded to the present sea level, are preserved in the lower 1.5 km of the creek course (Fig. 6.35) and have probably been formed as the river profile has adjusted to its present base level. Sea level stabilised about 6,500 years ago (Gibb 1986) but downcutting has continued as a result of tectonic uplift and some profile shortening by coastal erosion. Terrace surfaces are therefore inferred to be younger than about 6,000 years. An appreciation of this connection between sea level change and fold formation provides a powerful tool for understanding fluvial chronology. In the next section I examine deformation in the region of the Kate Anticline uplift and show that the events induced by the sea level changes and recent tectonic uplift can be correlated.

6.5.5 GEOMORPHIC EVIDENCE OF DEFORMATION

The coastal stream is a good example of the effects of profile shortening, as a result of sea level fluctuations throughout the Quaternary. These changes, which induced downcutting over the total length of the present profile are indicated by the migration upstream of sharp knickpoints.

Figure(6.40) shows the present profile and an estimated undistorted former profile for the Kate Creek and its associated uplifted coastal plain (M1). River bed heights, indicated by asterisks, are taken directly from the latest edition (1973) of the

NZMS 1 topographic map (enlarged to 1:25 000, Plate 3) using the digitizer. Measurements of the uplifted coastal plain height (Tiromoana coastal plain M1) were also taken from the same topographic map and these are shown by open circles.

To determine the amount that a terrace surface has been lifted, it is necessary to know its height relative to the riverbed when it was formed. Figure (6.40) illustrates the method of estimating the minimum differential uplift within the deformed zone, on the assumption that an undisturbed river profile grading to the former coastal plain (M1) can be used to model the original terrace profile. On both profiles a dotted line is drawn to indicate the shape of the undisturbed river profile.

An obvious convexity marked by a sharp knickpoint at 400 ft occurs on the youngest terrace surface (i.e. the Kate swamp) upstream of the axis of uplift. This knickpoint is a migrating knickpoint which is adjusting the creek profile to present sea level. The presence of an actively degrading creek in a narrow meandering valley is indicative of degradation both upstream and downstream of the deformed zone.

The lower section of the swamp surface is uplifted approximately along the axis of the Teviotdale Syncline. These geographical relationships can be seen in Figure (6.40) and the geomorphic map (Plate 11).

Comparing the amounts of vertical deformation of the lower swamp surface, and the estimated undistorted river bed profile to the age of their formation provides a means of estimating contemporary rates of uplift in this area (Fig. 6.40). There is no evidence for determining the ages of terraces on the Kate Creek valley, other than that from

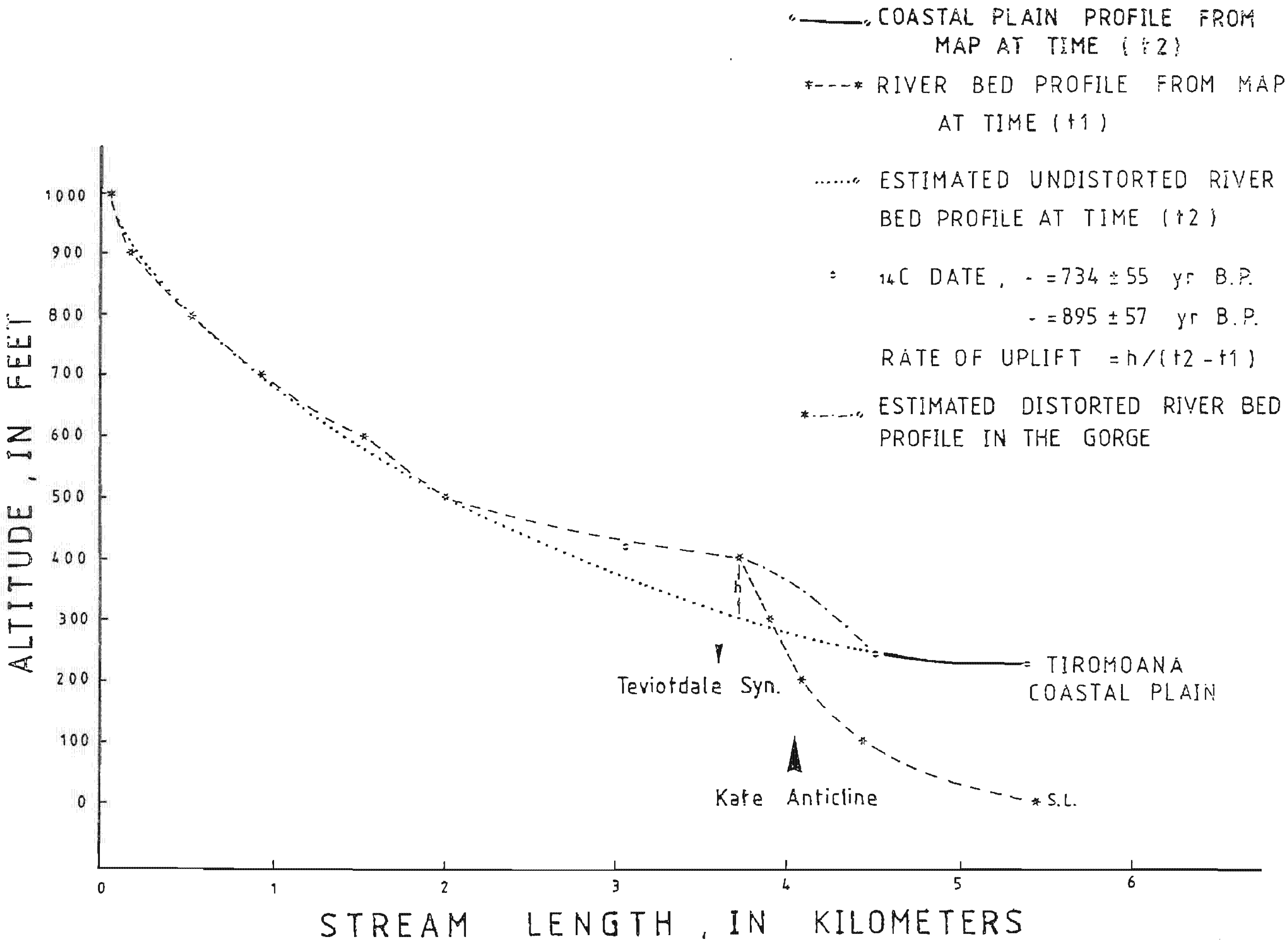


Figure 6.40 Method of estimating minimum rate of differential uplift due to
a zone of upward deformation

the swampy surface collected by the author, which give ages of 734 +/-55 yr B.P. for the upper surface, and 895 +/- 57 yr B.P. for the lower surface. If it is assumed that the age of the Tiromoana coastal plain (M1) is 60 ky, then the average uplift rate for the Teviotdale Syncline since that time can be expressed by the equation h / (t2 -tl). Therefore, the average rate of uplift is estimated at 0.51 mm/yr.

Growing folds are defined by differences in uplift values, the crests of anticlines having the highest values, and the troughs of synclines the lowest (Ghani 1978). The location of the axis of maximum uplift for late Quaternary growth of the Kate Anticline is east of the knickpoint (Fig. 6.40). This suggests that the location of maximum uplift may have affected the marine terraces on the southern limb of the Kate Anticline. Careful examination of the Last Interglacial marine terraces reveals that they have been broadly warped north of the Dovedale Stream mouth area into an anticlinal structure plunging northeast, by the longitudinal extension of the Kate Anticline propagating across the area (Plate 1).

Uplift rate for marine terraces can be calculated by taking the difference between the present height of the former shoreline and its original formation height and dividing by the age of the terrace to produce an average rate of uplift in mm/yr. If the Tiromoana coastal plain (M1) was formed c.60,000 years ago, (-21 m) must be subtracted from the present height to correct for its formation at a lower eustatic level. The uplifted Tiromoana coastal plain (M1) is a typical marine terrace bordering the seaward side of the Kate Anticline, where the cut platform surface is exposed beneath the terrace deposits at c.68 m above sea level (S68/220078), hence the uplift rate since that time has averaged (68 - (-21))/60 = 1.48 m/1000 yrs or 1.48 mm/yr. If this average rate has been maintained throughout the period of formation of the marine terraces bordering the Kate Creek gorge and if the assumptions upon which this calculation is based are correct, the highest Leonard marine terrace found at a height c.183 m above sea level would have been formed about 123,648 years ago, i.e. the Last Interglacial sea level maximum. Thus, terrace 2, 3 and 4 correspond to the three well known high sea levels of the, last interglacial, at c.80, 105 and 125 Ky. The three major high sea level events are correlated to oxygen isotope stages 5a, 5c and 5e respectively (Shackleton & Matthews 1977). The validity of this method of estimating the age of the Leonard's surface is possibly contradicted by the evidence for both differential, and non-uniform uplift rates throughout the region presented in this study. The discrepancy between the projected age and the 125 Ky age of the nearest known sea level stand may be a measure the order of magnitude error induced by these secondary effects when superimposed on a regional uplift and averaged over periods in excess of 100,000 years.

This rate of uplift is comparable with other calculated rates of uplift throughout the study area - namely an uplift of 1.61 to 1.79 mm/yr for the axis of the Black Anticline. Quantitative comparison of these uplift figures must be considered carefully. The uplift of the Kate Anticline derived from marine terrace data incorporates the sum of both regional and local differential uplift. The internal warping of the Kate Syncline derived above represents local differential movement only. The data from the Black Anticline is similarly based on reconstructed river profiles and is not a measure of total uplift.

6.5.6 EVIDENCE OF THE EFFECTS OF TECTONISM ON STREAM MORPHOLOGY

The sinuosity values of the Teviotdale River and Kate Creek where they cross the projected traces of actively growing folds were measured (Figs. 6.41 & 6.42). For each plot the location of the proposed uplift axis and the channel pattern are shown.

From these plots it is apparent that both streams are relatively straight along the axial region of the actively forming syncline which controlled the alignment of the streams. The courses become highly sinuous at the upstream side of the uplift axis. For example, the Kate Creek course becomes highly sinuous at the downstream end of the swampy area occupying the core of the Teviotdale Syncline, where the structure closes round the southwesterly plunging axis. The course changes abruptly and straightens where it cuts across the anticline, with deep incision marked by a sharp knickpoint. It is quite possible that the creek has modified its course and pattern in order to increase the slope in an area where tectonism is reducing it. In addition a detailed examination of the geomorphic features of this area reveals that a deviation of 50 degrees southward of the mean flow direction is quite evident (Plate 11).

The Teviotdale River is the most complex of all those studied, because it undergoes many changes in pattern along its 9.7 km course (Fig. 6.42). Since the uplift of the Kate Anticline the migration of the upper meander course has undoubtedly been inhibited. The river, in a sense, has been wedged between the topographically high homoclinal structure to the north and the growing Kate Anticline to the south (Stereomodel 19). Immediately after emerging from its upper meandering valley, it is only weakly sinuous (S = 1.2), and is

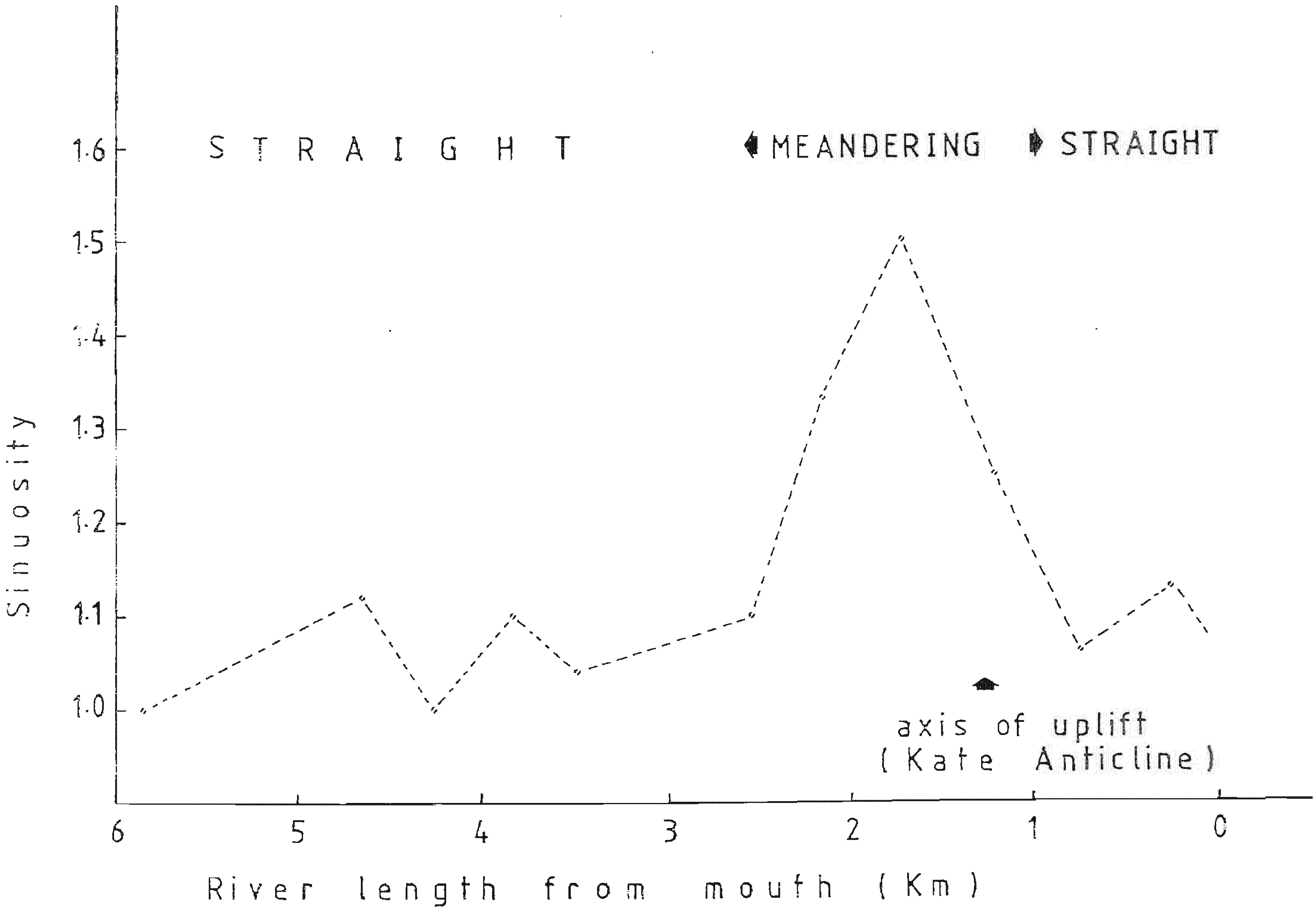


Figure 6.41 Relationship between river sinuosity, axis of uplift and distance

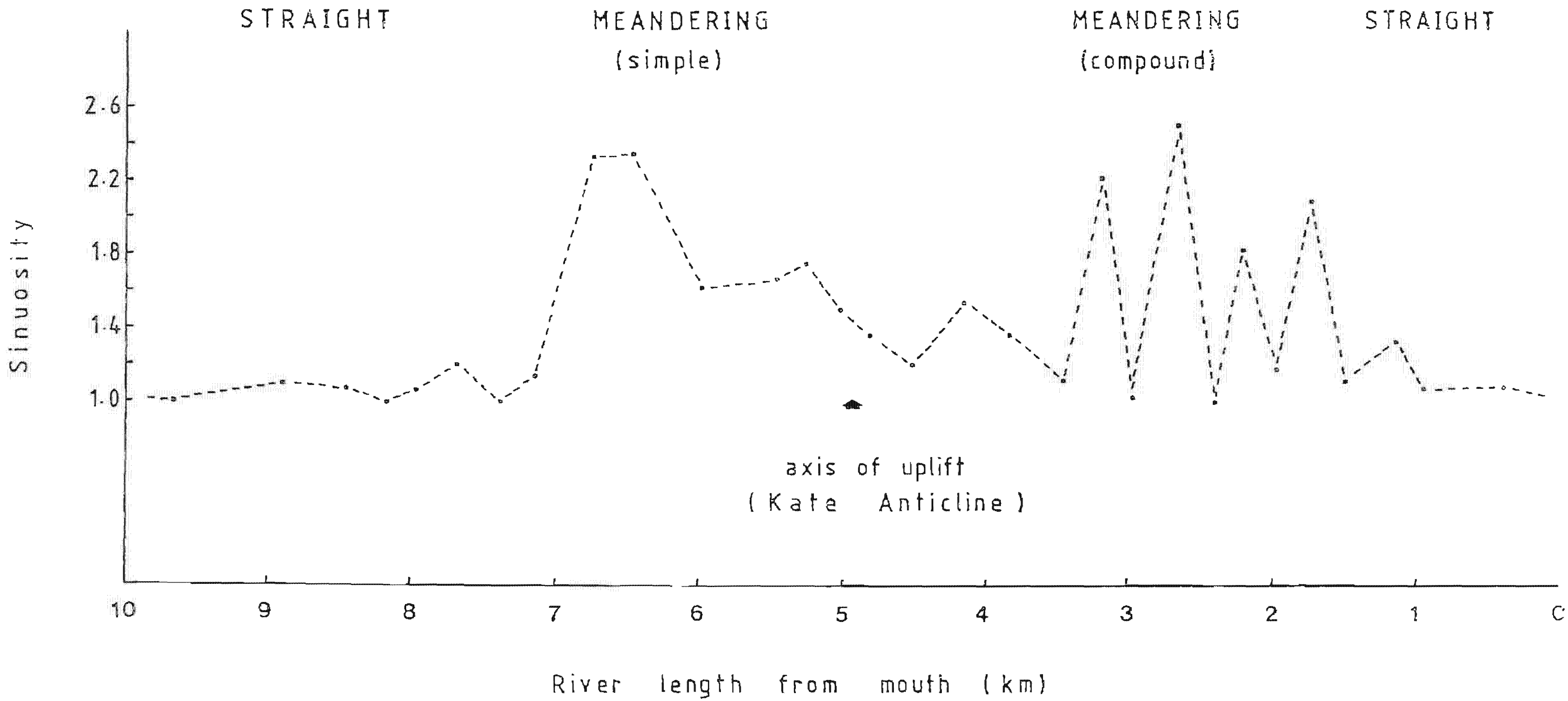


Figure 6.42 Relationship between river sinuosity, axis of uplift and distance of the Teviotdale River.

slightly incised in the valley bottom. Further downstream it shows a complex meander pattern deeply incised into the broad ancestral Waipara River valley. Field evidence indicates that the river pattern here is controlled by a resistant tilted bed of the Tokama Siltstone (Fig. 6.45). Thus, the channel displays markedly straight charactersitics parallel to the strike of the resistant bed, but alternates with highly sinuous reaches (S = 2.5) where it follows the bed dip direction (see Plate 11).

The morphology of the meandering valley of the Teviotdale River parallels the Waipara on a smaller scale. For example, an increase in the sinuosity on the upstream side of the Kate Anticline (Fig. 6.43) is accompanied by an upward convexity in erosional terraces over the projected trace of the crest (Fig. 6.44). There is also a tendency for slip-off to occur in the downplunge direction. Upstream aggradation occurs along the axis of the Teviotdale Syncline (plate 11).

6.5.7 CONCLUSIONS

Geomorphic modification of the Teviotdale River and Kate Creek have evolved in close association with active folding. The Kate Anticline at the edge of the shear zone was folded in late Pleistocene times and is one of the most actively growing anticlines in the study area. Movement or movements appear to have occurred in spatially irregular pulses probably related to intermittent fault movement at depth through late Quaternary time. This young anticline has warped deposits ranging in age from 125,000 yrs B.P., the time of the Last Interglacial marine terrace (M4), to at least 700 yrs B.P., the time of the Kate swamp surface.

Figure 6.43
Landforms along the Teviotdale River gorge. View looking north. Note the incised meander valley and flights of erosional terraces.

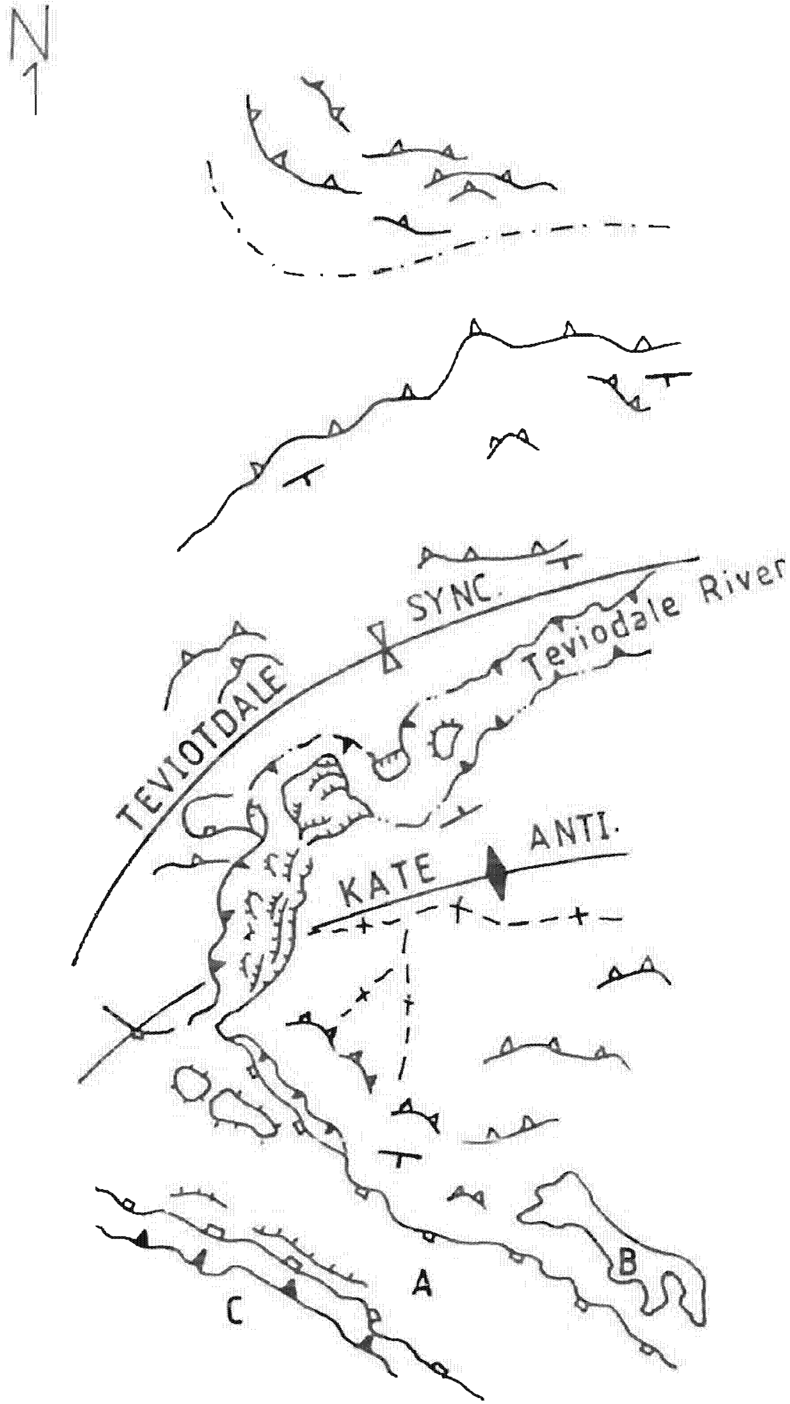
Figure 6.44
View of an upward convexity in erosional terraces (arrow), looking south.

Figure 6.45
Exposure of tilted, conglomeratic bed in the lower Teviotdale River gorge. Note the river course follows the strike of the resistant bed.

Geomorphic analysis, using continuity of landform, elevation height, tilt, and degree of preservation, indicates that the earliest established movement of the Kate Anticline appears to predate the highest marine terrace (M4). Active growth of the Kate Anticline resulted in differential tilting of the three marine terraces which are well preserved along the southern limb. The axial trace of Kate Anticline is sigmoidal and the coast line is oblique to the general trend of the fold. This means that gradients vary from place to place, generally increasing inversely with the distance from the anticlinal axis.

Stereomodel 19

Aerial view of Teviotdale gorge, cut through the southern end of the Kate Anticline. Of particular interest is the antecedent gorge. A direct correlation between structure and topography can be seen. Note the ancestral Waipara valley (area A) incised below the upper aggradation surface (area C), and the uplifted marine terraces (area B).


N
TEVIOTDALE
SYNC.
Teviodale River
KATE
ANTI.
A
B
C

6.6 THE MOTUNAU RIVER

6.6.1 SETTING

The Motunau River and its tributaries form the second major drainage basin in the study area. It rises on the eastern slopes of the Black and Cass Anticlines, at c.425 m elevation, flows eastward across the Montserrat Syncline to Glendhu, then southeast to Motunau beach (Fig. 6.46). The length of the main channel is 17.2 km, and drains a total area of about 68.5 km^2.

Uplift and river length reduction by sea level change are considered to be the major factors which have led to the rapid downcutting by many rivers in the coastal range of the Waipara region. Preserved older terrace surfaces in the lower Motunau valley indicate that the river, which is now deeply entrenched in a narrow meandering channel up to 50 m deep, was originally wider (Fig. 6.48). Terracing by the Motunau River has resulted in the removal of large portions of the Motunau coastal plain. These terraces which are considered to post date the Motunau alluviation (Carr 1970) are separated from the plain itself by well defined and distinctively cut scarps.

Carr (1970) has suggested that the platform irregularities may possibly have been caused by gentle warping of the Motunau coastal plain, and he refers to the high and low anomalous areas as the Motunau Anticline and Syncline respectively. According to him at least one, the synclinal axis, appeared to have been active during the shore platform cutting phase. The anticlinal axis had tilted the Motunau estuarine beds and therefore the latest movement was taken to have been Oturian (?Otiran) in age (Carr 1970). In addition, he demonstrated that the old high aggradational level of the Motunau

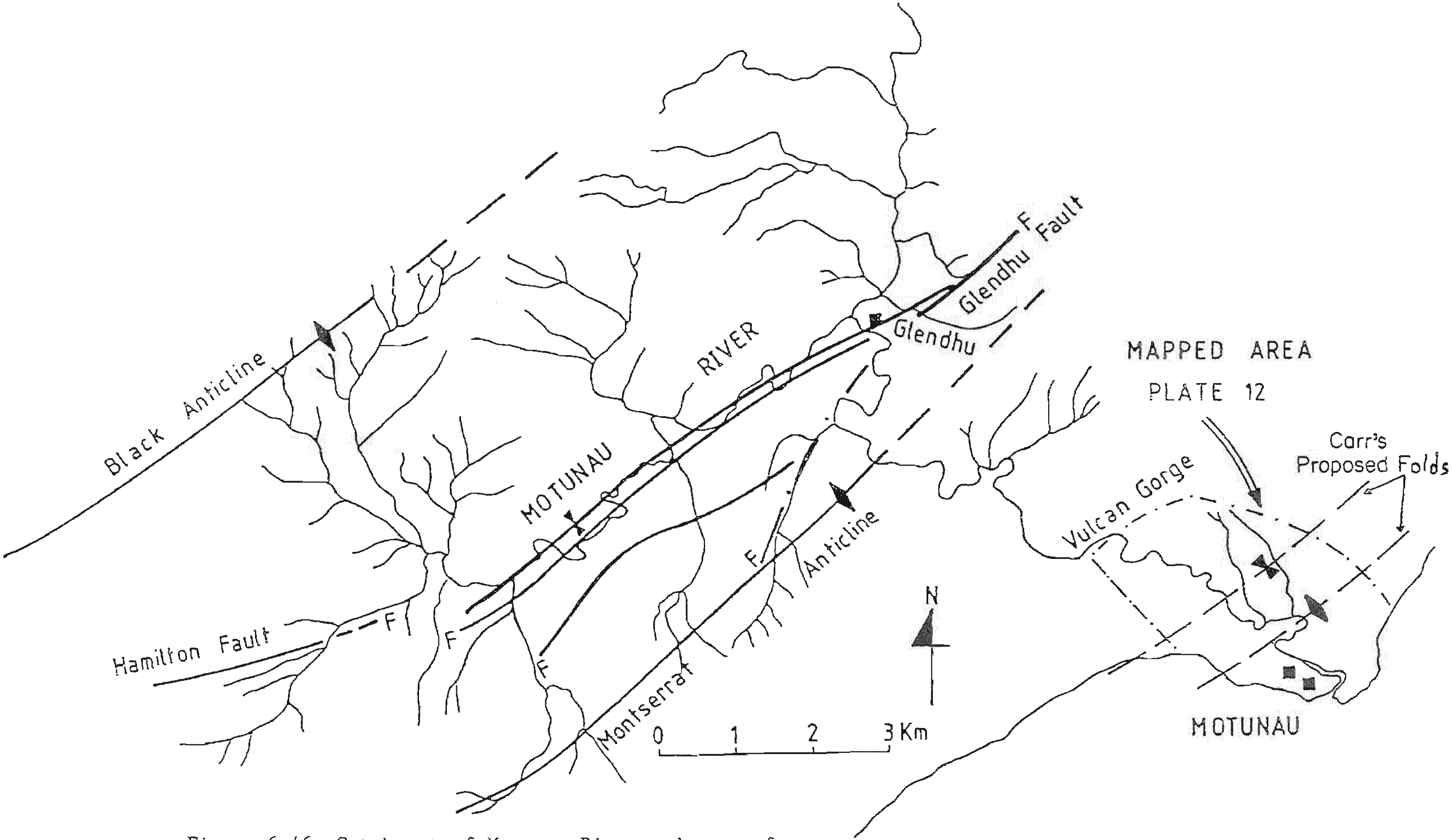


Figure 6.46 Catchment of Moyunau River and area of river and marine terraces mapped for this study.

River is referred to the Motunau coastal plain largely on morphological grounds, and that it has been warped and tilted to a degree similar to that of the plain. River degradation followed or was coeval with the terminal regression of the sea subsequent to estuarine deposition. All post-Motunau terraces are unwarped (Carr 1970, p.184). Aspects of this interpretation are considered critically below.

Initially Carr's structural interpretation was accepted and on noting the strongly developed incised meander pattern, this area was chosen as a detailed case study. This is because it was thought to illustrate the early stages of the kind of antecedent processes already described in a more evolved form for the other catchments discussed already in this chapter. Thus this study had two related objectives. The first was to determine the main response of a meandering river to anticlinal uplift and synclinal subsidence on the major and minor folds, and the second objective was to study the denudational chronology of the lower Motunau River valley with particular emphasis on the influence of more subtle structures and of climatic control upon its fluvial geomorphology. In the event the existence of Carr's Motunau folds is questioned and the meander pattern attributed to a different cause.

Pertinent to the question of the degree of possible deformation and shear, is the paleomagnetic data of Walcott et.al (1980). Within the limits of data collected in a section from Tongaporutuan to Kapitean sediments in the lower Valcan Gorge, they are unable to show any significant departure of the direction of magnetization from the present time. Mean declination is given as 354^{o} with a standard error of 21.7° based on 36 sample sites.

A geomorphological map was prepared at scale of 1:10,000, using

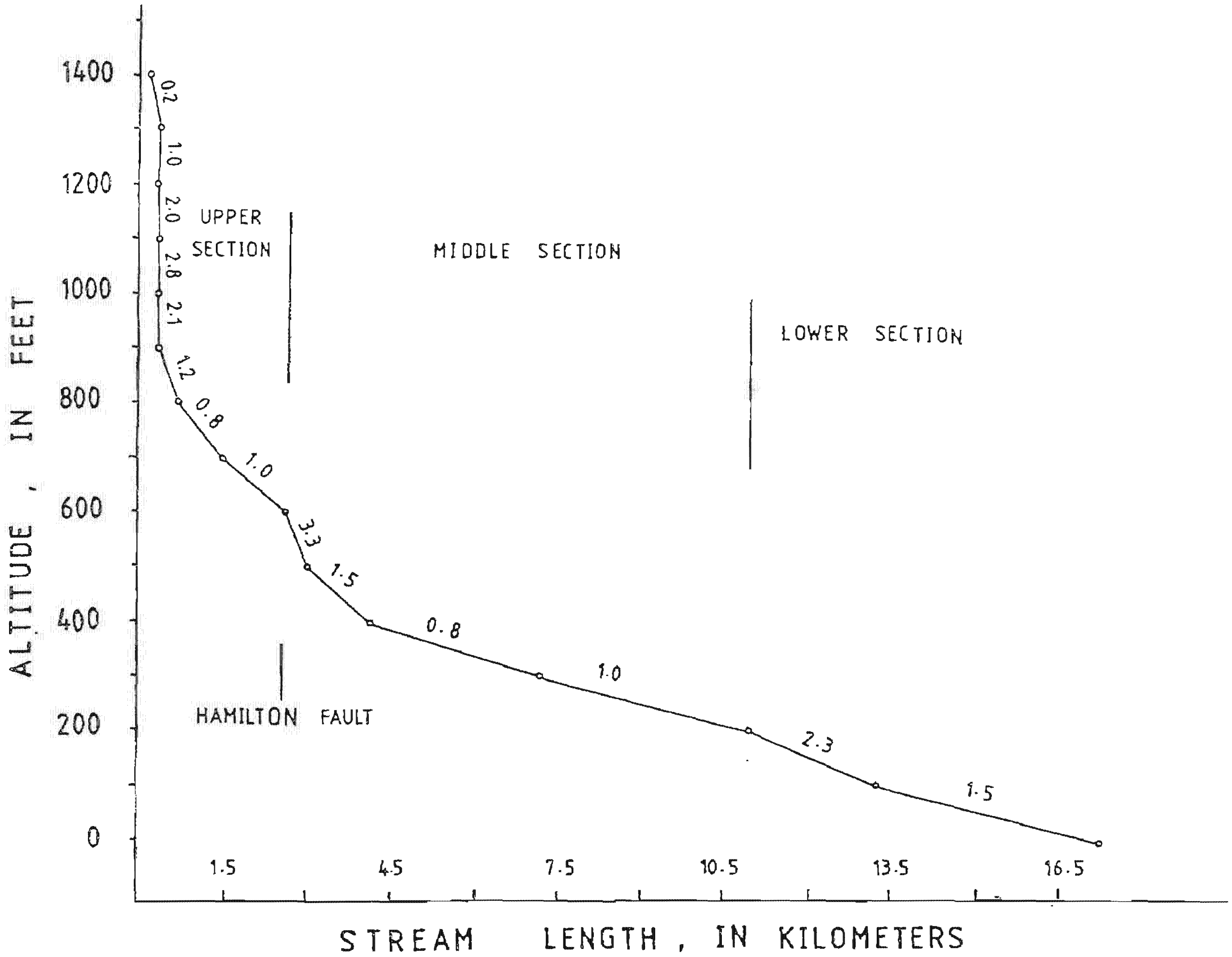


Figure 6.47 Longitudinal profile of the Motunau River. Numbers on river represent gradient index (GI/K).

enlarged aerial photographs (Plate 12). Surfaces were grouped according to their origin (marine, fluvial, structural and denudational) and age. These surfaces can be subdivided into smaller units using geomorphic and relief amplitude controls. The result is an interpretative geomorphological map which shows the distribution and origin of several geomorphic processes. Identification of fluvial and marine deposits was based solely on field character. Spot heights were taken from Carr (1970) data.

6.6.2 THE LONGITUDINAL PROFILE OF THE MOTUNAU RIVER

The present longitudinal profile of the Motunau River from the headwaters to the sea has an overall concavity that reflects a progressive decrease in gradient (Fig. 6.47), and it has three distinct sections. The Motunau River exhibits knickpoints around an altitude of 1100 ft, 600 ft and 200 ft.

The first knickpoint at 1100 ft corresponds to a change in lithology from the resistant limestone of the Omihi Formation to more easily eroded Oligocene sandstone of the Waikari Formation. The river presently cuts into bedrock for all but the lower 1 km of its upper section upstream from the Hamilton Fault (see also plates 2 and 4).

The second knickpoint at 600 ft, can be ascribed to the northeast extension of the Hamilton Fault, see Chapter 3, section 3.5.1, with indices that increase from 1.0 to 3.3 GI/K towards the fault.

In its middle section this river flows through a strike valley that has developed along a synclinal axis (Montserrat Syncline) and the associated parallel trace of the Glendhu Fault. The long profile of the middle section lacks abrupt changes in slope. Movements along the Glendu Fault have offset both upper terrace and the underlying bedrock whilst the valley floor was aggrading in the actively growing syncline (S68/314170). The features on this fault are amongst the least well preserved in the region and its latest displacement may be considerably older than latest displacement of most of the moderately active faults in the region, see Chapter 3, section 3.5.3.

At present, the Motunau River still retains a well-defined meander trace along a synclinal valley. In this section, not only is the thalweg incised within the floodplain, but the narrow, meandering floodplain itself is incised within lower terraces (see Plate 2).

The third knickpoint at 200 ft probably represents a knickpoint which is migrating as the river profile adjusts to present sea level (see Plate 4). This moderately high index (GI/K = 2.3) and the presence of an actively degrading stream in a narrow floored meandering valley are indicative of degradation. Thus, the knickpoint has developed in this case from a eustatic lowering of the base level accompanied by differential uplift. Rejuvenation by uplift is evident in the deep incision of the present lower section drainage below the uplifted marine terraces (Plate 12). It also appears that a rather sudden increase in sinuosity value in the vicinity of the Motunau coastal plain immediately downstream of the east dipping side of the Montserrat Anticline. Likewise, higher sinuosity values correspond to the lower valley slopes found upvalley of the uplift axis (i.e. Montserrat Anticline), but the lower Motunau River appears to represent a system which has evolved some differences which will be discussed shortly.

6.6.3 THE TERRACES OF THE MOTUNAU RIVER

The area of detailed study in the Motunau Catchment is the terrace development on the lower Motunau River (Fig. 6.48).

No previous detailed study has been made of these terraces. Carr, in conjunction with his work on the stratigraphy of the Motunau coastal plain, labelled 3 of this series of terraces, lying on both sides of the river terraces A, B and C Carr 1970; but did not establish a wider correlation.

Locally derived alluvial gravel and the bedrock which it overlies have both been cut by fluvial erosion which has probably been induced by both tectonic uplift and sea level change. Fans sloping eastward from the eastern flank of the Montserrat Anticline, grade to the coastal plain north of the Motunau River (Plate 12).

East of the Vulcan gorge, a sequence of seven erosional and two aggradational terraces which predate the modern Motunau beach has been identified below the level of the uplifted coastal plain (M2) (Plate 12). Terraces are numbered 1-9; the higher the number the older the age. Samples suitable for "absolute" dating are absent from both fluvial and marine deposits. Therefore, estimates of age and the correlation of terraces are made on the basis of the degree of preservation and the extent of terrace surface within the whole study area. If the inferred age of the uplifted coastal plain (M2) is correct (Fig. 6.52), then all the fluvial terraces are less than 80 000 years B.P. and equivalent to the oxygen isotope Substage, 5a.

A. Upper Aggradation Surface

The oldest and highest aggradation surface found in the lower Motunau River valley (terrace no.9) only exists east of the Vulcan gorge and again near the river mouth (Plate 12). All evidence of this

Figure 6.48

Terraces of the lower Motunau River cut into upper Tertiary rocks. The highest and oldest terrace, in the foreground, is the 80 ky marine terrace.

Figure 6.49

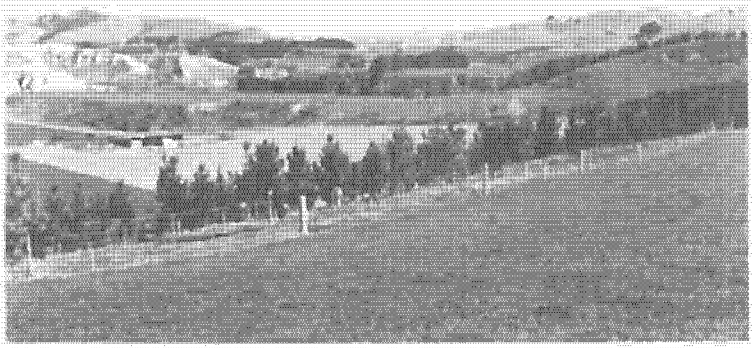
Close up view of fluvial deposits, illustrating an upper aggradation terrace, south bank of Motunau River near Vulcan Gorge.

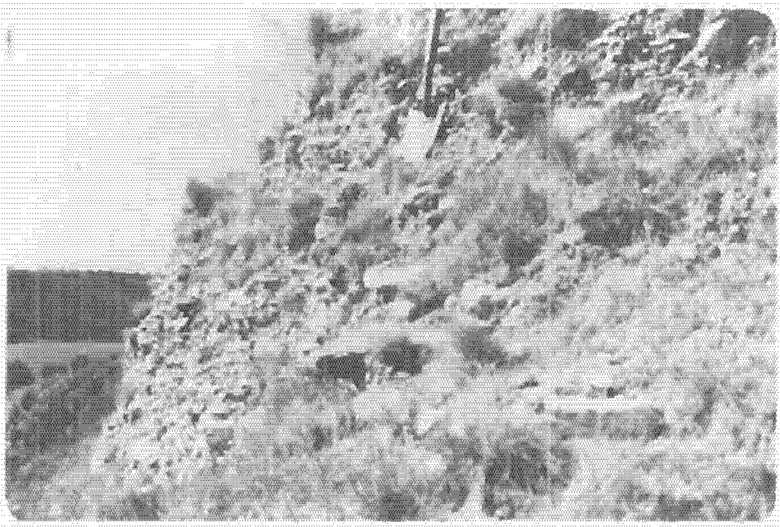
Figure 6.50

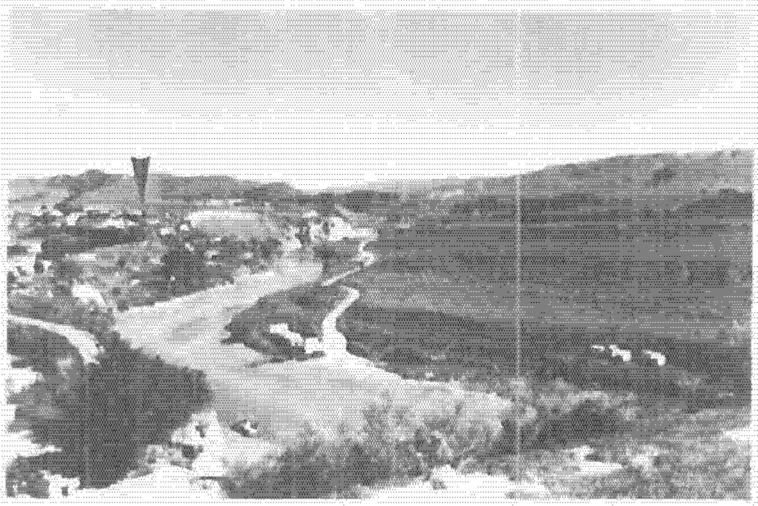
Close up view of fluvial deposits, illustrating a lower aggradation terrace. Note the strath surface (arrow) and the river deposits which become finer in texture towards the surface.

Figure 6.51

View looking northwest across Motunau River mouth. Note the high topographic surface (arrow) and the lower flight terraces.







surface between these two areas has been removed by a period of significant erosion across the valley. There are good exposures of terrace cover both upstream and downstream along the river banks. The terrace deposits consist of muddy angular gravel composed of blocks of limestone and Tertiary rock, angular greywacke pebbles, reworked greywacke beach gravels, mud and sand (Fig. 6.49). Terrace materials are nearly 12 m thick. About 0.5 km downstream from the Vulcan gorge, the upper aggradation deposits stand at 47 m above sea level and have a strath surface at 35 m. Near the river mouth the upper aggradation deposits are at 25 m. Altitude of the terrace surface above the cliff is 28 m.

B. Lower Aggradation Surface

The lowest and youngest aggradation surface found along the lower Motunau River is much wider in extent than the older terraces (terrace no.6, Plate 12). On the basis of underlying deposits and morphological characteristics this surface can traced from the Vulcan gorge to the sea and is recognized as evidence of a second phase of aggradation. Terrace materials are nearly 4 m thick and consist of about 3 m of fluvial gravels covered by 1 m of yellow-brown sand (Fig. 6.50).

C. Degradation Terraces

Two distinct periods of degradation terraces both above and below the lower aggradation surface are present in the lower Motunau valley (Fig. 6.51). They have a similar distribution to the aggradation surfaces, but these terraces are probably degradational, an origin which is indicated by the thinness of deposits underlying the terrace surfaces. Fine gravel and sand were observed to maximum depth of 1.5 m and many terraces are bedrock strath.

A sequence of five erosional terraces (terraces nos. 1-5) has been identified below the level of the lower aggradation surface. Two other erosional terraces (terrace nos. 7 and 8) have been identified below the upper aggradation surface.

6.6.4 GEOMORPHIC ANALYSIS

The tectonic deformation of rivers results in terraces that have diverging-converging patterns where fold axes are crossed or end abruptly at fault zones (Hull 1985). Geomorphic and tectonic investigations in this area strongly indicate that the anomalously high topography of the upland area near the Motunau Settlement (Fig. 6.51) is the product of differential erosion by ancestral coastal channels which are unrelated to tectonism. Detailed analysis of the geomorphic map, the terraces, dip and strike measurements, and subsurface data do not indicate the presence of either the Motunau Anticline or Syncline in this area (Plate 12).

The present Motunau coastal slope, which has a southeasterly dip, reverses its direction and rises anomalously in the vicinity of the Motunau Settlement. Field evidence indicates that the reconstruction of the higher fluvial deposits exposed under the upland area, are related to ancient drainage patterns. This period of deposition was followed by a period of significant erosion across the coastal plain.

Vestiges of an ancient drainage pattern that differs from the modern one remain. For example, exposures along the road to the Motunau Settlement, Fig. 6.55 , show a broad channel, cut some 4.5 m into the Greta Formation, which has been infilled with alluvial gravels and massive muds. The bottom of the scour stands at 35 m above sea level.

The capture of the old coastal channel by the continuous coastal cliffing is clearly seen in the geomorphic map (Plate 12). Near the large meander cut-off (S68/393144), the broad, U-shaped ancestral coastal channel once traversed the upland area now captured by the coastal ciffing. Note that the old high fluvial deposit should not be confused with the upper aggradation surface (terrace no.9), exposed both upstream and downstream of the topographically higher area, Carr appears to have done (Carr 1970).

Plate 12 shows part of the uplifted coastal plain (see Fig. 6.52) which is interrupted by numerous irregular steps of fluvial origin which are considered to have been formed early in the history of the plain's emergence and which have eroded a considerable amount of the uplifted surface, These erosional escarpments, seldom exceeding 6-10 m in height, do not demonstrate enough continuity to be able to reconstruct the ancient drainage network, although they are considered to have been formed by ancestral drainage patterns (plate 12). Thus, the meander gorge and its tributary valleys which are impressed upon the plain as a result of the onset of a second cycle of erosion caused by falling baselevel, due to differential uplift and sea level change, are post 80 Ky. The same factors have been responsible for major and minor gulley erosion of the coastal plain. This tendency towards strong downcutting is exemplified by the unnamed creek immediately north of Motunau River, and the captured creek to the south (Plate 12).

Tectonic tilting of originally horizontal and inclined surfaces

Figure 6.52

The 80 Ky marine terrace, Motunau coastal plain, showing thin beach deposits over a cut surface of tilted Greta Formation. Motunau Island in distance.

Figure 6.53

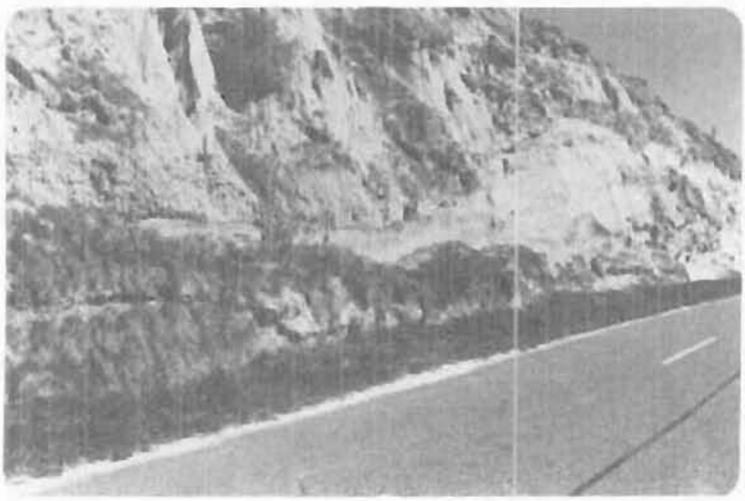
Exposure of resistant tilted bed of Mt Brown Formation which had affected the sinuosity of the lower Motunau River.

Figure 6.54

Close up view of upper Tertiary synsedimentary sliding (Mt Brown Formation).

Figure 6.55

Channel and fill in Greta Formation, exposed in road-cutting, near Motunau Beach settlement.



appears to have superimposed a seaward dip upon the upper Tertiary sequence. The cut platform (Fig. 6.52) originally sloped seawards, and probably towards the Motunau River mouth. Along the eastern margin of the Montserrat Anticline, a decrease in dip with increasing distance from the fold axis is shown (Plate 12). In addition, local variations in both the amount and the direction of the dip across the cut platform have been observed and are probably related to synsedimentary slumping. In general; the Montserrat Anticline appears to have behaved in a similer way to the Cass Anticline.

Field evidence suggests that a rather sudden increase in sinuousity occurs in the vicinity of the old synsedim en tary slumping on
has maintaned
the north bank of Motunau River (Fig. 6.54, S68/394152). This is a common occurence when a river, as a result of incision, encounters resistant materials in a portion of its course (Fig. 6.53). It becomes more sinuous and then straightens again once the barrier has been crossed. Here such a resistant barrier, in this case bedrock, has caused the compression and deformation of the upstream meanders (Plate 12). The Motunau River thus shows similarities to the pattern displayed by the Teviotdale River downstream from its upper gorge.

The meandering lower course of the Motunau River have maintained very high sinuousity values near the river mouth. Meander loops have grown and two meanders cut-offs were found (Plate 12). Channel migration and meander patterns appear to have typified events in the post-glacial period in this valley.

6.6.5 CONCLUSIONS

On the basis of detailed field studies and geologic and geomorphic maps it seems well established that the Motunau Anticline and Syncline are primary platform irregularities rather than true tectonic features. However, certain other platform irregularities near Boundary Creek to the north of the study area are clearly of tectonic origin (see Plate 5). Here folding and faulting appear to have affected former shorelines as described in Chapter 5, section 5.8.

The lack of evidences for shear (en echelon) folding as far as the coast is consistent with the paleomagnetic data.

All of the available evidence indicates that the topographically high area adjacent to the Motunau Settlement formed as a combined result of the topographic relief due to differential erosion, and the subsequent deposition of significant amounts of sands. There is no need to invoke a tectonic explanation for the development of the high area, nor is there sufficient evidence to support such an explanation.

The height, geomorphic position, and the thick fluvial deposits underlying these terraces indicate that the upper aggradation surface (i.e. terrace no. 9) is a likely correlative of the Teviotdale Surface in the lower Waipara mouth area. Furthermore, the lower aggradation surface (i.e. terrace no. 6) is a little lower than the Motunau estuarine beds at the river mouth and is the likely correlative of the Canterbury Surface.

The Motunau River provides a useful illustration of the care that needs to be taken in using parameters such as sinuosity and terrace irregularities to identify sites of active deformation.

6.7 THE RAKAIA RIVER

6.7.1 INTRODUCTION

The Rakaia valley is an area in which steep topographic and climatic gradients combine to produce a range of dynamic geomorphic and sedimentary environments quite different from those of the coastal range of the Waipara region. The Rakaia River has the highest mean flow of all Canterbury rivers.

The braided river pattern of the major inland valleys is typical of an aggrading riverbed, forming wide, flat valley floors. The Rakaia River is a straight, steep, active, braided river throughout most of its course but it has a short mid-section, which meanders through a hard-rock gorge (Plate 14).

The behaviour of the Rakaia River at the point where it crosses the up-thrown side of a reverse fault demon strates anomalies in the valley and channel characteristics. For these reasons, this area was chosen to compare and contrast with the landscape and deformation styles of the Waipara region.

Two distinctive physiographical regions are recognized. The upper catchment is mountainous and extends from the headwaters near the Main Divide downstream to the Rakaia Gorge and covers an area of 2,640 km^2. The lower catchment is a narrow riverbed corridor which extends 60 km from the sea to the Rakaia Gorge (Plate 14).

The study area dealt with in this section is that immediately surrounding the gorge through which the Rakaia River passes between these two physiographic regions, before issuing from the mountainous region of the Southern Alps on to the Canterbury Plains (Plate 13).

The study discusses how faulting and climatic changes influence geomorphology, and speculates, from data gathered in the Waipara region, about the origin of typical features.

6.7.2 SETTING

The Rakaia Catchment covers an area of 2,910 km^2 and is defined by the surface catchment boundaries of the Rakaia River and its tributaries (Plate 14). The river is about 150 km long and extends in a south-easterly direction from the Main Divide to its mouth on the east coast of the South Island.

Three principal lines of evidence support a history of antecedence across an actively developing fault.

1. From the Main Divide to the sea, the Rakaia River crosses only one structural barrier in an antecedent gorge and the bed form changes from braided reaches of aggraded flood plain to a deeply incised single channel in the bedrock (Fig. 6.57). In the Rakaia Gorge, the transition into the up-thrown (backtilted) side of the reverse fault block is marked by an abrupt change to an anomalous, incised, meander pattern which straightens again downstream of the fault trace (Fig. 6.59). The occurrence of these meanders interrupts the normal pattern of the Rakaia River.
2. The Rakaia Gorge is one of the most impressive examples of late to post-glacial erosion. It has been incised below older fluvioglacial outwash and underlying Tertiary rocks leaving a series of erosional and aggradational terraces (Figs. 6.56 & 6.58). The narrow gorge which is only about 125 m wide has almost vertical, bedrock walls rising 50 to 130 m above the present river, and represents a distinct

Figure 6.56

Rakaia River terraces in the Rakaia Gorge, ranging from the recent low-level terraces to the elevated and backtilted Aranuian terraces.

Figure 6.57

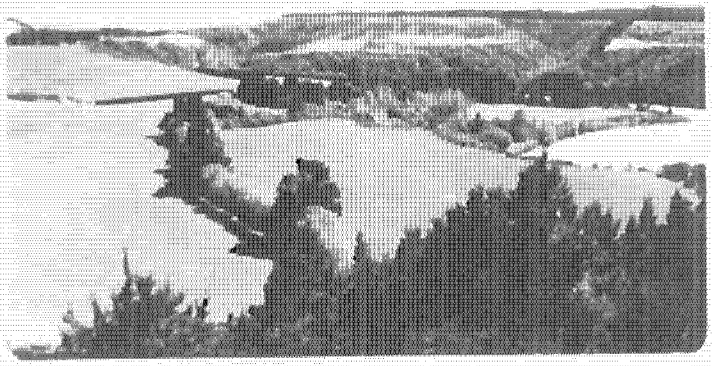
The Rakaia gorge showing a deeply incised single channel in bedrock. The bedrock walls rise 50 to 130 m above the present river.

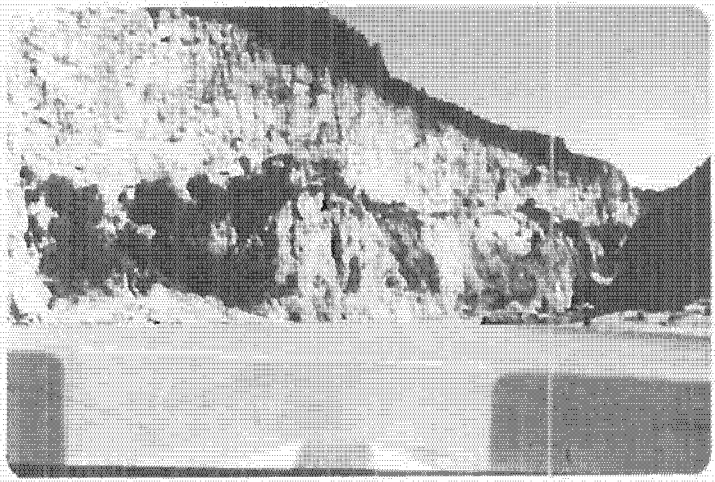
Figure 6.58

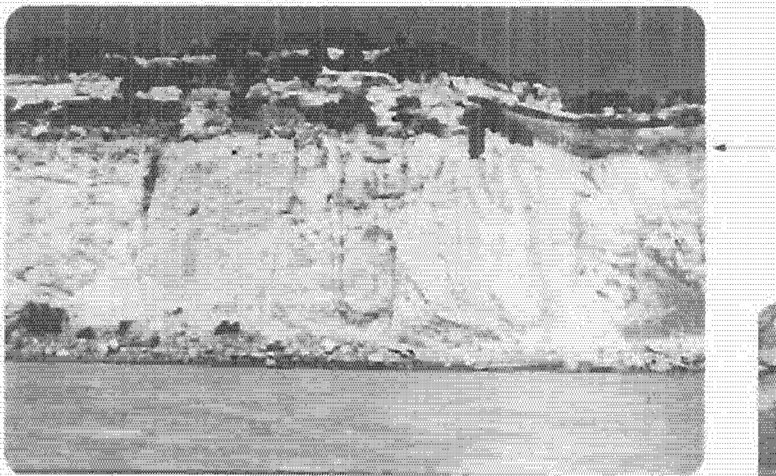
Typical river terraces through the upper Rakaia Gorge. The present strath surface (arrow) exposed in the valley sides.

Figure 6.59

The Rakaia River cutting through the Canterbury plains. The valley is slightly over 1.5 km wide. Note the braided river system.







topographic departure from the broader Rakaia valley both upstream and downstream.

3. An interesting feature of the morphology of the gorge is the evidence it provides of the influence of tectonic control upon the form of the course of the river and its tributaries. The backtilt of the strath surface and the diversion of nearby Boundary Stream appear to be strongly dependent upon the repeated movement of this fault.

6.7.3 THE GEOLOGY OF THE RAKAIA CATCHMENT

The stratigraphy may be divided into three major units; the older basement rocks, the covering strata and the younger glacial and fluvial deposits. Plate 14 is a simplified geological map showing the geographical distribution of the main rock types.

The basement rocks, which form the ranges of the Rakaia Catchment, consist mainly of pre-Cretaceous, moderately to strongly indurated sandstones, "greywacke" and alternating sandstone/mudstone sequences. In the Whitcombe Pass region, these sediments have been metamorphosed to a low-rank schist. Minor rock types such as conglomerate, red/green siltstones and jaspillites occur at several localities within the catchment. Terrestrial calc-alkaline volcanic outcrops of the mid-Cretaceous age are found in the Rakaia Gorge and also form the Rockwood and Highpeak Ranges to the northeast of the gorge (Oliver 1977, 1980).

The covering strata consist of a succession of late Cretaceous to early Pleistocene coal-measures, greensands, limestones, siltstones and sandstones together with a range of both intrusive and extrusive igneous rock, which lie unconformably on the basement rocks. Only scattered remnants of the covering strata are exposed within the catchment.

Late Pleistocene glacial deposits, fluvioglacial outwash gravel and postglacial alluvium fill the valley floors and form a part of the Canterbury Plains. A detailed description of the late Pleistocene deposits and their relationship to each ice advance and retreat for part of the Rakaia Catchment is given by Soons (1963), Soons & Gullentops (1973) and Carryer (1967). The Waimunga Glaciation is represented by the Woodlands Advance, and the Otira Glaciation by the three later advances - Tui Creek, Bayfield and Acheron.

The Canterbury Plains portion of the Rakaia Catchment is a flat to gently undulating alluvial plain ranging in altitude from sea level to about 490 m a.s.l. near the foothills. Late Quaternary glaciers extending down the Rakaia Valley (Plate 14), and the coalescing fans of late Quaternary outwash gravels have greatly contributed to the formation of the Plains.

6.7.4 THE EFFECT OF TECTONIC STRUCTURE ON THE RAKAIA CATCHMENT

Deformation of the basement by folding and faulting during two major orogenies is a controlling factor in the present day drainage pattern. Fault zones of severely crushed rock form lines of weakness with a low resistance to erosion by rivers or ice. Viewed from a distance (i.e. by Landsat imagery, see Chapter Seven), the rectangular drainage pattern, resulting from faults oriented predominantly NW/SE or NNE/SSW, becomes apparent. Some of the observed fault lines are shown on the geological map (Plate 14) while probable faults, now buried by glacial and postglacial deposits, are believed to control the course of the Upper Rakaia River and its tributaries.

Evidence of this structural control remains today. For example, the Harper Valley follows the Harper Fault which possibly continues in the upper Rakaia River above Manuka Point.

The pre-glacial drainage pattern and relief produced by the Kaikoura Orogeny has been modifed by late Pleistocene glacial erosion. Glaciers modified the pre-glacial drainage pattern by rounding sharp bends into curves and truncating the ends of spurs with the overall effect of straightening the river valleys. The valleys were deepened and widened, and their slopes were steepened.

The Rakaia River has constantly adjusted its profiles across the glaciated valleys in response to a combination of events including tectonic uplift along the ranges (in contrast to the tectonic stability or slow subsidence of the rest of the Canterbury Plains), sea level and bed load changes.

The Rakaia River has become antecedent across an upward movement of the late Tertiary bedrock which now forms the gorge inlier near the S.H.72 Gorge Bridge (Fig. 6.60), indicating ongoing development of an irregular faulted surface beneath the Plains.

6.7.5 THE GEOMORPHIC MAP OF THE RAKAIA GORGE

A geomorphological map was prepared using enlarged aerial photographs and field data at scale 1:10,000 approx. (Plate 13). Landforms related to glacial, fluvial, structural and denudational origins were recognized in the study area. These forms and the processes involved in their formation are clearly indicated by line symbols and as such these units are self-explanatory. The morphogenetic landform units are indicated by different colours. Further subdivision is made on the basis of lithology, geomorphological processes, and relief amplitude.

The banks of the Rakaia Valley illustrate the impor tance of constructional landforms such as glacial, fluvial and debris fans in preserving evidence of recent landscape history.

A. Units of Glacial Origin

The effects of glaciation are most spectacular in mountainous terrain. Stereoscopic photographs of the Rakaia Gorge reveal erosional and depositional landforms with great clarity, and most of them are unlikely to be misinterpreted on aerial photographs. The glacial landforms were marked on aerial photographs the features attributed to each advance were drawn on the geomorphic map (Plate 13).

Correlation of the glacial deposits mapped in this area was made using geomorphological evidence from the sequences of previous workers. Largely on the basis of the field relationships of these deposits, Soons recognised four major ice advances in this part of the valley and named them The Woodlands, Tui Creek, Bayfield and Acheron Advances. Each advance was followed by a retreat phase, during which, the deposits of the previous periods were dissected as the regime of the river changed in response to altered climatic conditions and the ice-front retreated into a depression previously occupied by the glacier. Subdivision of the advances has been based largely on the presence of moraines whose outwash surfaces are clearly distinguishable from each other. It is not the purpose of this study to contribute to the study of glacial deposits or to consider chronological subdivisions. The primary purpose of this research is both to produce a more detailed geomorphological map than has hitherto been published, and to examine the behaviour of the Rakaia River and contrast it with the dominant modes of deformation and landscape formation found elsewhere in the study area.

i. The Woodland Advance

The Woodland Advance (Soons 1963, Soons & Gullentops 1973, Carryer 1967) is the earliest and most extensive glacial advance recognised in the Rakaia Valley although deposits from it are more discontinuous. Within the area of this study, the Woodlands Advance is represented by a flat terrace about 1.5 km in length, at an altitude of 820 m on the valley wall above Blackford Homestead (Plate 13). Correlation of this deposit with the Woodlands Advance is based on the position of this surface above all other glacial deposits in this part of the valley.

ii. The Tui Creek Advance

Moraines and outwash surfaces of two advances, Tui Creek II and III, are shown on Plate 13 extending southward from Pinedale Homestead. The vertical separation is about 20 m (Soons & Gullentops 1973). The remnants of the Tui Creek I moraine, almost completely buried by Tui Creek II outwash, occur 5 km down the valley (S82/202553).

iii. The Bayfield Advance

The Bayfield Advance has been divided into two minor advances, (Soons 1963, Soons & Gullentops 1973). Deposits of these advances (Plate 13) are separated from those of the Tui Creek Advances by a clear difference in height, which may be traced on both sides of the

Rakaia River for some distance down the valley from the area of the end of the moraines near Cleardale Homestead. They are 30-60 m lower than the Tui Creek moraines and outwash surfaces. Carryer (1967) recognised a further advance (the Blackford) preceding Bayfield I and two additional minor advances within Bayfield I Advance. A recent study by Soons and Burrows (1978) of the Lyndon-Acheron Valley, which links the Rakaia and Waimakariri drainage systems reexamined the stratigraphy and showed that the Bayfield Advance is represented by three advances, indicated by well-separated outwash surfaces. Plant materials in the silts and in overlying peat stratigraphically separating the two younger advances have been dated. The basal silts gave an age of 22,200 +/- 750 years B.P. (i.e. Bayfield II minimum age) and the peat 19,200 +/-550 years B.P. (i.e. Bayfield III maximum age Soons and Burrows 1978).

In view of the complexity of the Bayfield Advances and these new dates, the Bayfield III Advance is considered in this study to be appropriately grouped with the Bayfield II Advance.

vi. The Acheron Advance

The Acheron Advance, is the fourth ice advance described by Soons (Soons 1963, Soons & Gullentops 1973), and includes all advances in the middle Rakaia Valley later than the Bayfield, and was divided into three minor advances.

Within the area of this study, the Acheron I is represented by a complex of moraine ridges near Cleardale Homestead. These deposits do not show any weathering. They form a thin layer on top of the earlier deposits, which they have not extensively disrupted. No outwash surface is associated with the Acheron I moraines (Plate 13).

The evidence for the existence of lakes in the Rakaia Valley is widespread, and has been the subject of discussion in a number of studies (Cox 1926, Lauder 1962, Soons 1963, Soons & Gullentops 1973). At least once, a lake was formed between the retreating ice and the deposits of its advance, as is indicated by lacustrine silts in the valley and by remnants of lake shorelines.

A series of lake shorelines near Cleardale Homestead (Plate 13) records three levels of a post-Acheron I lake at present altitudes of approximately 420, 405, and 380 m a.s.l (Soons & Gullentops 1973).

The dating of the organic remains in the silts near the silt/gravel transition in a section of the terrace face (S74/073729), has been given as 11,650 +/- 200 years B.P. (Burrows 1979).

B. Units of Fluvial Origin

i. River terraces

Plate 13 indicates the arrangement of the drainage system of the study area. The modern Rakaia River is flanked on both sides by an extensive series of aggraded and degraded terraces (Fig. 6.58). Above and below the gorge these terraces are cut out of the glacial deposits and covering strata and may be of considerable length. In the vicinity of the gorge these terraces are, in general, much more restricted in extent and are more complex in arrangement than those above and below the gorge (Plate 13).

A sequence of eleven erosional terraces has been identified later than 19,200 +/- 500 year B.P. but predating the active floodplains. Terraces within this age group form the majority of those preserved in the Rakaia Gorge and are typified by the presence of river gravels, or bedrock strath and identified by their sequential relationship (Figs. 6.58 & 6.65). Above the inner gorge, the relatively broad U-shaped channel of the old Rakaia River is clearly seen in profile. Sediment is supplied to the river at multitudinous locations along its channel by bank and bed erosion, rockfall, rockslide and debris avalanching.

ii. Flood plains

Two floodplains are distinguished on the modern Rakaia River both upstream and downstream from the Gorge, whereas in contrast, only point bars are recognised in the narrow meandering reaches (Plate 13).

iii. Alluvial fans

A detailed study of the alluvial fans on the south bank of the Rakaia valley has shown that their truncation by either glacial action or, more often, by river action has usually resulted in renewed fan building. Renewal of debris supply from the headwaters of the streams is the other main factor causing rebuilding of the fans. The fans vary considerably in morphology and extent as a result of the characteristics of the catchment area and the local erosion base. The profiles of the fans are moderately convex and some fans are composite, displaying both fluvial and debris flow constuction (Plate 13).

C. Units of Denudational Origin

The southern parts of the study area are characterized by the high mountainous terrain of the Mt Hutt Range. The basement rocks of the area form the foothills of the Southern Alps and rise abruptly to a height of about 2000 m above the Canterbury Plains.

The northern flank of the Mt Hutt Range illustrates the importance of constructional landforms such as alluvial, colluvial and debris fans in preserving evidence of recent landscape history. The effect of rock type and structure on rock and slope stability for the northern flank of the Mt Hutt Range has been described by Carryer 1967.

Excellent examples of rockfall, debris flow and gully and rill erosion can be seen along the cliffs of the Rakaia Gorge (Plate 13).

E. Units of Structural Origin

On both banks of the Rakaia Gorge, the isolated remnant of an uplifted fault scarp of rhyolite, pitchstone and andesite can be clearly identified in the field by their form and resistance. A recent study of these volcanics by Oliver (1984) showed that the rhyolite and andesite were, on the basis of chemical analysis, petrographic and paleomagnetic studies, and K/Ar dating, correlatives of those in the Mt Somers area of Mid-Cretaceous age. The stratigraphic relations between the two volcanic types has been in dispute ever since Haast (1872), but the K/Ar dates (Adams & Oliver 1979) give an age range of approximately 87 to 95 Ma for the andesite and rhyolite.

The Rakaia Bridge Fault (Fig. 6.61) was known from ground studies prior to this air photo investigation, although its full extent was unknown. The rest of the fault following the east-northeast trend was revealed for the first time by this study.

Locally the geology is very variable (Plate 13), with the topographically overlying andesite flows probably in fault contact

Figure 6.60

Rakaia Bridge Fault (arrow) and terraces, Rakaia Gorge Bridge section. View looking southwest from the left river bank. Note the bed-rock terraces and the uplifted volcanic rocks on the up-thrown side of the fault. For more details see Figures below.

Figure 6.61

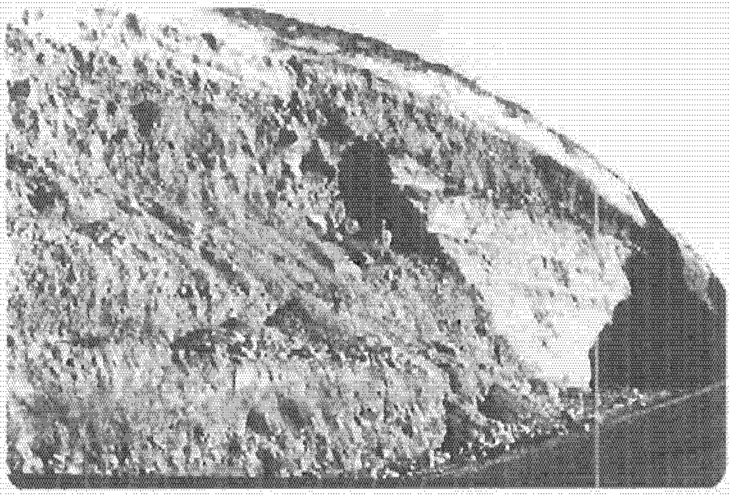
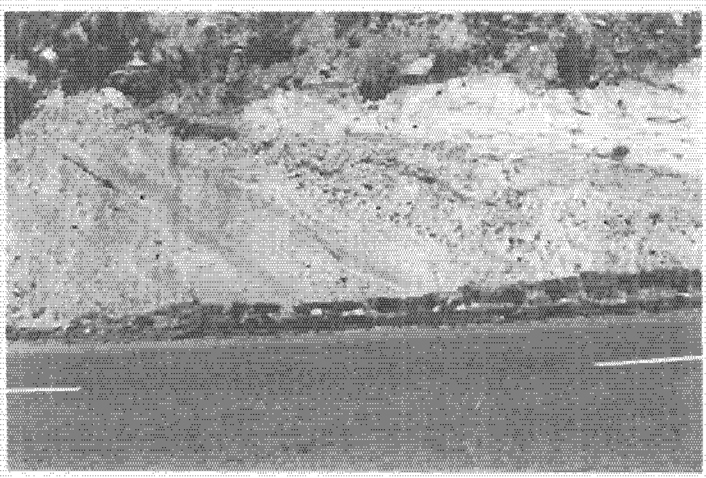
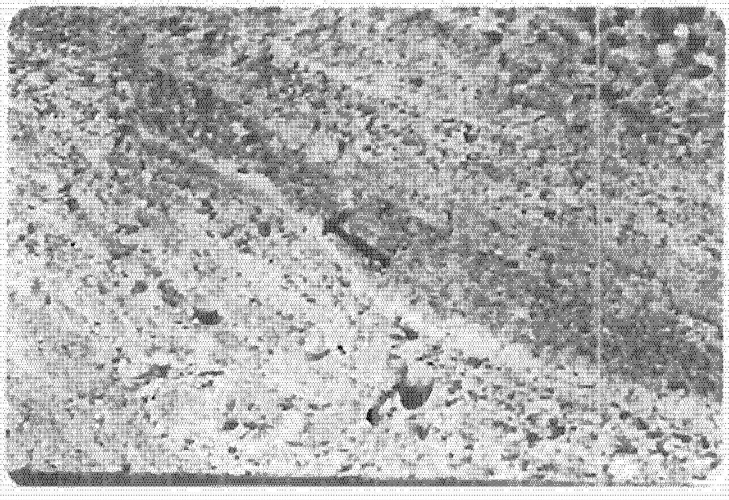
Close up view of the reverse fault in deposits in the Rakaia Gorge Bridge section. Note the fault movement continued until a relatively recent date, but prior to deposition of at least the uppermost part of the loess cover (photo: J.M. Soons).

Figure 6.62

Deposits in the Rakaia Bridge section downstream from the fault trace. An angular unconformity can be seen between each unit. This was caused by synchronous fault movement and sediment depcsition.

Figure 6.63

Close up view of the fine greywacke gravel (arrowed unit in Fig. 62), yellow-brown weathered, well stratified. Tilted up-stream at 4-5°.

with the rhyolite beneath Round Top and with the basement rock of the Mt Hutt Range. This range, which rises abruptly to a height of about 2000 m above the Canterbury Plains is probably bounded by south-westward extension of this fault.

Late Quaternary movements probably offset a small channel in the vicinity of Mt Hutt Ski Road (S82/063555) Plate 13. The channel now follows a fault trace. The existence of a probable fault scarp and an old channel can be used as secondary evidence for vertical movement along a fault, its relationship to the fault giving the sense of displacement, which is opposite, in this case, to the sense of movement at the gorge.

Although little work has been done along this fault it is thought to play an important role in the deformation zone that exists within the eastern foothills of the Southern Alps. If this is true, it supports the present interpretation of the Rakaia Bridge Fault. The Rakaia Bridge Fault is part of a northeast-southwest trending zone of Quaternary high angle reverse faulting which runs along the eastern side of the Southern Alps. However detailed mapping is needed to prove this hypothesis.

Active deformation of the Rakaia valley is known, particularly at the northern bank of the Gorge. Here late Quaternary faulting is apparent in the form of well marked fault-traces, of which one of the most striking is the "Rail road" fault (Plate 13) near the Gorge (Speight & Dobson 1924). Toward the northwest end of the fault substantial, fluvio-glacial deposits are faulted as shown on (Fig. 6.64). No definite evidence of displacement of the beds on the north bank of the river is, however, available.

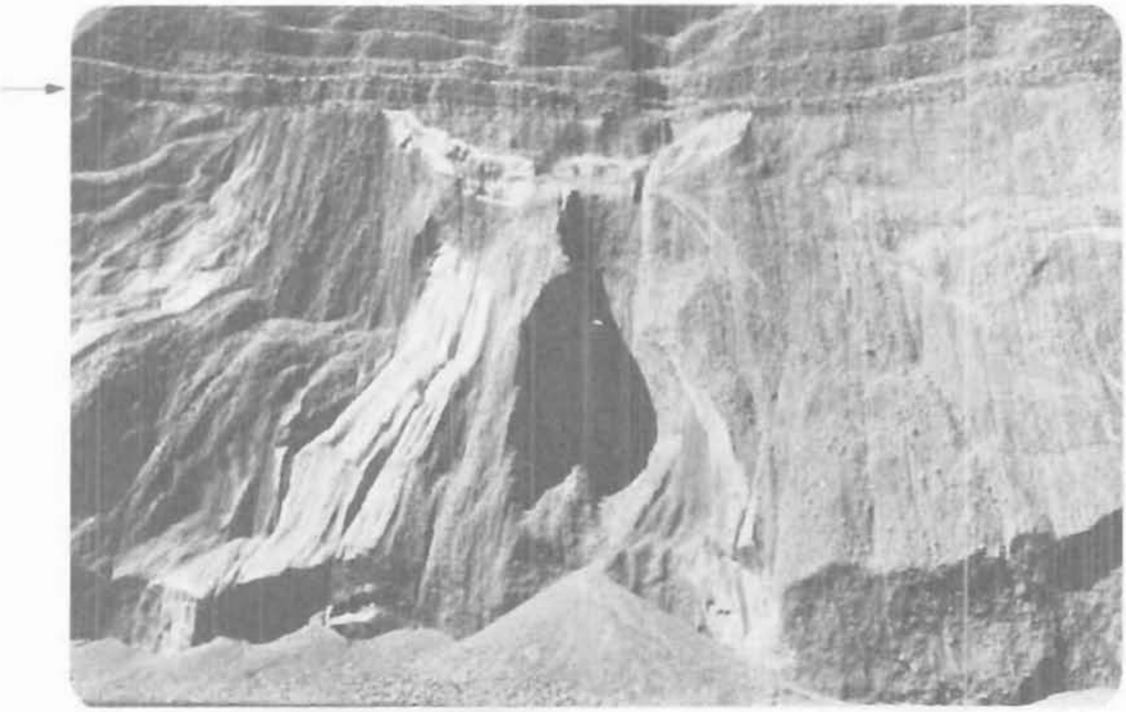
A sequence of late Tertiary sediments unconformably overlies the

Figure 6.64

Well delineated fault zone ("Rail-road fault") exposed along the northern bank of the Rakaia River. No evidence of displacement of the strath surface (arrow) is observed.

Figure 6.65

In this photo the strath surface (arrow) shows a well-defined backtilt on the upthrown side of the Rakaia Bridge Fault.



volcanic rocks. The coal-measures overlying the rhyolites and andesites in the Rakaia Gorge are the only members of this Tertiary sediments exposed in this area (Cox 1926). They cropout as part of a basin-shaped fold which is tilted towards the south-west, so that half of the basin is obscured below the Pleistocene deposits on the south side of the river. The maximum thickness exposed is about 300 m.

6.7.6 GEOLOGICAL HISTORY OF THE RAKAIA GORGE

The Rakaia Gorge is the major physiographic feature of the Rakaia valley. The magnificent terraces and the peculiar situation of the gorge have attracted studies of its geology and geomorphology.

Two explanations have been put forward concerning the history of the Rakaia Gorge. Cox (1926, p.108) specifically states:

"...The presence of Pleistocene silts and conglomerates above the gorge to a depth of at least 500 ft below the level of the top of the barrier of resistant rocks now forming the gorge inlier, requires that either (1) the bed of the river before the deposition of these beds was eroded to that depth, and therefore the barrier cannot have occupied its present position; or (2) this basin in the older rocks represents a trough hollowed out by the glacier which at one time occupied the valley, and whose deposits form the base of the Pleistocene series".

Lauder (1962) describes the valley as the "Rakaia Graben" on the basis of stratigraphic evidence. It has been suggested (Speight 1933) that the Rakaia Valley may owe its major lineaments to faulting. Although the straight flanks of Mt Hutt and the Big Ben Range overlooking the valley offer some support for this, no evidence as to the actual lines of any such faults has so far been presented.

The overall form of the Rakaia valley has been well described by Soons 1963 and Soons & Gullentops 1973. The framework of the valley is aggradation and degradation of the glaciated valley into the floor of which the river has re-entrenched itself, and has landform detail reflecting modification by a variety of denudational processes (Plate 13).

The detailed geomorphic history of the Rakaia Gorge remains to be studied. Some recent re-examination of the factors that control channel patterns of alluvial rivers of the Canterbury Plains by Carson (1980 a & b) shows the concept of a "threshold" to have limited validity, for a number of reasons. In particular the concept of a threshold level of stream power or shear stress for the onset of braiding, is essentially empirical and offers little in terms of a mechanistic understanding of the problem.

Carson (1980 a) has also indicated that meandering and braiding may have existed in concert with long-term degradation of the channel bed throughout much of the Holocene. He remarks that:

"... the Rakaia River upstream of the gorge, took the form of a drift-blocked lake. Relatively rapid dissection of the drift-filled gorge area, together with aggradation in the downstream part of the Rakaia sandur, a glacially-enlarged trough upvalley of the gorge, has merged the profiles of the upper and lower reaches, following the major glaciation responsible for outwash on the Plains".

There is a widespread practice of referring to braiding and meandering as alternative forms of stream behaviour. As the rivers progressively add to their bed-material loads downstream, incision gradually gives way to a shoaled channel, and meandering is replaced by braiding. A recent study by Rundle (1985) of the Rakaia River, demonstrated that a major influence in the formation of multiple channels is the mechanism of diversion and piracy by the headward extension of chutes at low water-levels across a tongue-shaped bar formed at higher water levels.

Changes of valley-floor gradient provide another explanation of downstream pattern variations. A stream crossing the trace of a reverse fault should be affected by such slope changes in ways similar to the effects of decreased gradients upstream of the axis in an actively growing anticline as has been discussed before (see section 6.1.2). In response to such slope deformation, the stream could have degraded and aggraded its channels and also changed its channel patterns (Fig. 6.2).

But, if braiding is to occur, degradation must be sufficiently slow to produce wide, shallow channel floodplains accessible to dissection, and bed sediment loads must be sufficiently high to ensure that occasional, local thalweg-shoaling will force flood flows out of the main channel. Experimental studies by Ouchi (1985) and others show that the main features of braided-stream response to uplift (i.e. upward movement and backtilting) are (1) alternating bars with a braiding tendency in the upstream reach; (2) terracing and degradational trends in the central area; and (3) bar-braided patterning in the reach downstream of the uplift.

The initial fault movement created a disturbance of the slope of the active stream channel resulting in downcutting becoming greatest in the uplifted area. Channels on the backtilt slope are deeply incised in the river bed leaving terraces at a higher level as a result of changes in base level (Plate 13).

I agree substantially with the antecedent origin for the Rakaia Gorge.

In 1970, re-alignment of State Highway 72 on the western side of the Rakaia Gorge Bridge (S82/139585) clearly exposed deposits of the Woodlands and younger glacial advances, and were described by Soons and Gullentops (1973). Evidence of progressive fault movement can be seen where the cutting section is divided into three by two discontinuities, one a fault (Fig. 6.61) and the other a buried terrace face. Yellow-brown gravels and silts at the base of the eastern end of the section (Fig. 6.62) are correlated with those on the western side of the fault (Fig. 6.61), so that the throw on these beds is estimated to be about 40 m. Higher in the section, unconformably overlying terrace gravels are displaced less, the throw being about 1 m, while the capping loess is undisturbed. This has masked the fault scarp so that no trace is visible on the terrace surface, either on the ground or in aerial photographs.

Progressive movements may have little immediate effect on alluvial rivers such as the Rakaia, but the cumulative effect of changes in the slope of the alluvial valley can be great (Figs. 6.62 & 6.65), and this is most pronounced when the exposure of sediments of differing resistance in the bed of the channel causes a change of channel morphology (Fig. 6.60). Terraces of the Rakaia River have been cut deeply into the fluvioglacial outwash deposits and underlying Tertiary strata during the last glaciation. During interglacial and interstadial periods the continuing down-cutting must have produced gorges almost as deep as the present gorge. The most important problem raised, and not entirely resolved by the exposures along the road in the Rakaia Gorge area is the lack of absolute age controls on these terraces and the difficulty of correlating the deposits. Therefore, realistic estimates of the rates of fault movements are not possible at this stage.

Plate 13 shows that the Rakaia River has maintained very high sinuosity in its course above the fault trace during the Holocene. Meander loops have grown and been cut off in different locations several times, but they have never migrated past the fault trace. Downstream of the gorge, on the left bank of the present braided river, low terraces show clear signs of a meandering channel. A reasonable explanation of these meanderings is that they are surface expressions of vertical movement along an underlying fault close to the Rakaia Bridge.

The evidence of the Gorge Bridge section suggests that at least two periods of fault movement can be identified. For example, the yellow-brown deposits of an older outwash at the base of the section are tilted 4-5^{o} up-valley (Fig. 6.63), as are the rhyolite outcrops exposed at the lower end of the gorge. The stream gradient of an alluvial system would decrease in magnitude as the stream traversed a cross-cutting, upthrown fault block. The decrease in gradient would increase the number and amplitude of stream meanders, thus enhancing deposition of point-bar sands (Plate 13). In this meandering valley, the Rakaia River has adjusted laterally across the glaciated valley and has been able to cut down through it. Thus, the entrenchment of the Rakaia River into the older covering strata was later than 19,200 +/- 550 year B.P..

Detailed examination of the geomorphic features across the Rakaia Bridge Fault trace suggests that upward movement and backtilting, initiated the diversion of the ancestral Boundary Stream

(Plate 13) which could not keep pace with these changes.

The modern Boundary Stream is actively eating back into a high level terrace area on the north side of the Rakaia Gorge. Upstream of the fault trace, there is a well defined meandering pattern with a deeper channel and narrow floodplain. Downstream, the valley becomes more confined and passes through a narrow bedrock gorge of volcanic rocks, similar to those in the Rakaia Gorge, before opening out into a wide-floored reach leading to the confluence with the Rakaia River (Plate 13). The Boundary Stream thus shows a downstream progression of a pattern that is close to the sequence displayed by the Rakaia River but on a smaller scale.

6.7.7 CONCLUSIONS

The most important additions to our previous knowledge of the Rakaia Gorge area, which have resulted from the work outlined above, concern the antecedent origin of the gorge and the use of geomorphic features to document the tectonic history of actively growing faults. An attempt is therefore made in this study to explain the varying channel patterns of the Rakaia River as a response to changes in valley slope deformation, particularly those resulting from actively growing reverse faulting. It is evident that streams as contrasting in stream power as the Rakaia River and Boundary Stream are sensitive to the rates of uplift and tilt generated by active tectonic processes. These effects can still be identified even where flow regimes are as strongly affected by other disturbances to sediment and water volumes as take place during dynamic fluvio-glacial cycles.

CHAPTER SEVEN

LANDSAT IMAGERY ANALYSIS

7.1 INTRODUCTION

Landsat imagery has been a significant aid in extracting basic geological and geomorphological information from low and high relief terrains.

Photo-interpretation of Landsat imagery requires optimum presentation of the data in order to utilize fully the available information. Optimum presentation first requires careful image selection. This necessitates consideration of the fundamental climatic and physiographic features of the study area. This study examines an area of contrasting relief in the eastern South Island, New Zealand, from the Rakaia River in the south to the Marlborough shear zones in the north (Fig. 7.1). Topographically the area ranges in elevation from sea level to over 2200 meters. This area was chosen because of its varied geologic, tectonic, geomorphic and topographic nature. Comparision of the image with the geomorphic interpretation map (Plate 15) shows that topography is controlled by geologic structure.

The area consists of indurated "greywacke" type rocks of Upper Jurassic and Lower Cretaceous age (Torlesse Supergroup), as well as less indurated Cenozoic covering strata, unconsolidated alluvium and marine sediments of Quaternary age, and Holocene dunes and still forming floodplain deposits along the rivers.

The Landsat investigations were therefore made with the following objectives in mind:

firstly, to develop available methodology for the the rapid production of small-scale geological and geomorphological maps from

Landsat imagery;

secondly, to identify new geological features previously unrecognized at ground level and finally, the ultimate objective as a potential tool to delineate the evolution of drainage and landscape features associated with active tectonic folding and faulting from small-scale remotely sensed images.

The visual geological and geomorphological interpretation of Landsat images discussed in this chapter follows earlier field work supplementing Wilson's basic stratigraphy and structure of the coastal range of the Waipara region, so that the lithology and structural control on the geomorphology is well understood, and is supported by limited field analysis of selected areas outside the target area. Stereoscopic viewing instruments were utilized to assist with the visual interpretation.

7.1.1 LANDSAT IMAGES FOR THE CURRENT RESEARCH

The advantage of having information from several bands of energy becomes apparent in this study, and the use of drainage-texture as well as tone (colour) allows the identification of more features on the colour composite images. Sophisticated image processing, involving spatial filtering and multispectral manipulation of the data were applied only to the Landsat images E-2282-21254. These techniques can provide further information not retrievable from photographic images.

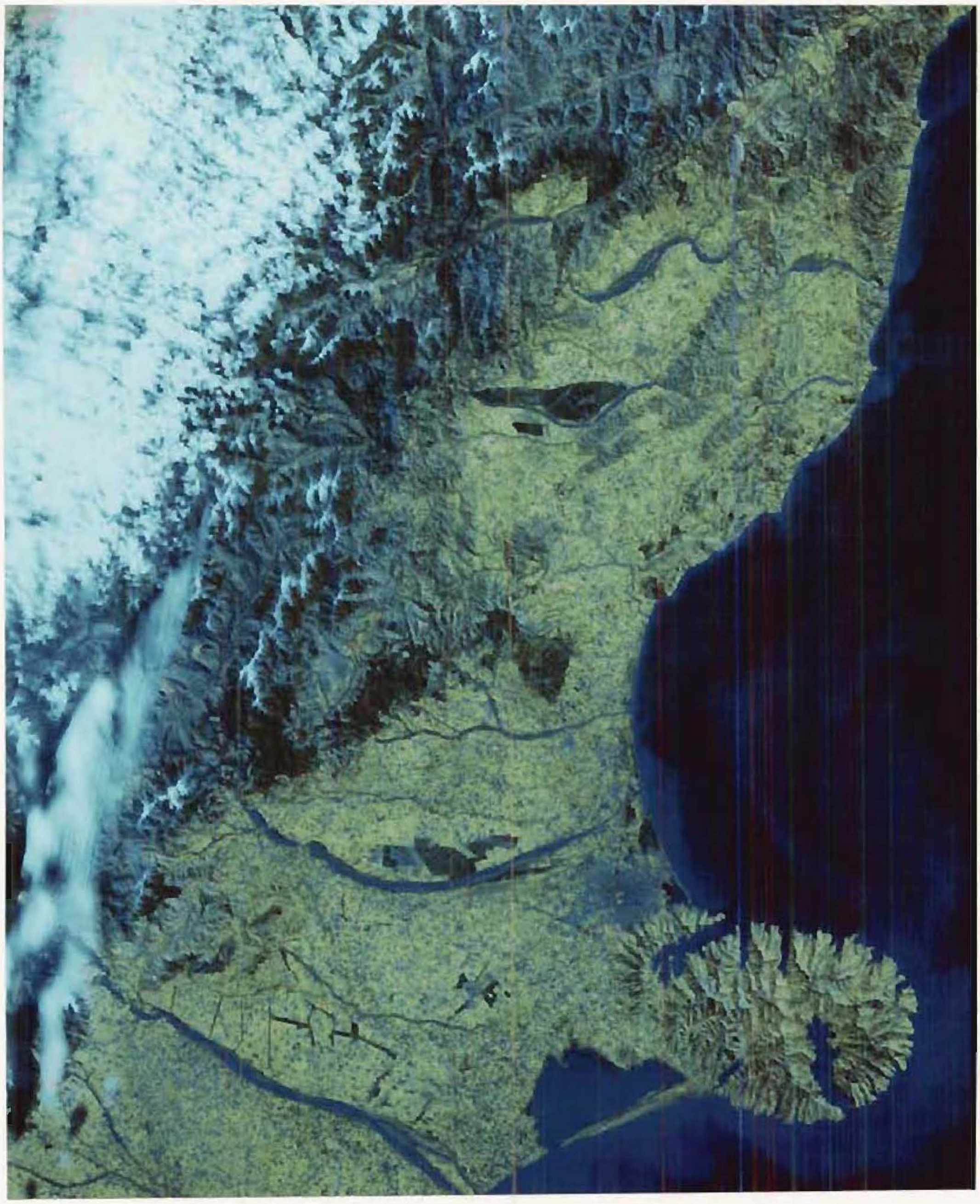
The effect of sun elevation and direction on the visibility of the lineaments can be demonstrated by a comparison of the images E-2192-21265, E-21272-21063, and E-2282-21254 (Appendix 4). The first two of these, has a sun elevation of 15^{0} and 09^{0}, and an azimuth of 046^{0} and 048^{0} respectively. This low sun angle in the winter images

Figure 7.1

Landsat image of the North Canterbury-South Marlborough region. Enhanced color composite Landsat 2 image of 2282-21254, aquired October 31, 1975 PEL 7.

Figure 7.2

A video processing system at PEL, Wellington. Input is provided by the television camera on the left. The analyst can directly control the degree of contrast stretching, density slicing and edge enhancement. Various types of processed images can be displayed on the monitor.



of the study area favoured the observation of lineaments in the lower relief terrain, but not in the high relief terrain of the Southern Alps. The image E-2282-21254 has a sun elevation of 43^{0} and an azimuth of 064^{0}. This image was recorded on 31 October 1975. Although it covers only the eastern division of the Southern Alps covered by the other two images, the relative number of lineaments is greatest in this 31 October image (Fig. 7.1). A colour enhanced image of MSS band 4 has been shown to highlight active faults; the enhancement apparently a consequence of the differential reflectivity in the fault crush zones as a result of cataclasis and possibly water seepage.

7.1.2 IMAGE DATA AND PROCESSING

The following imagery data were utilized in the work reported here:

1. Color composite of 2192-21265, acquired 2 August, 1975 PEL 13 Bands 1, 2, 4, Pos. Enh. using Cols7-2200 1.50 Range Eye Corrested, Landsat 2, scale 1:1,000,000.
2. Enhanced-colour composite Landsat 2 image of 2282-21254, acquired 31 October 1975 PEL 7. MSS 4 and principal components, 1, 3, 4 are coloured red, blue and green respectively. H.E. 2340-3264, 09:03:39, B1, scale 1:1,000,000.
3. Landsat RBV of the north Canterbury-south Marlborough region, acquired 6 October , 1978, P-N L 2 NASA Landsat E-30215-21282-B.
4. Landsat E-2192-21265 MSS Band 7; SUN EL 15, AZ 046, aquried 2 August, 1975, scale 1:1,000,000

5. Landsat E-2282-21254 MSS Band 5; SUN EL 43, AZ 064, acquired 31 October 1975, scale 1:1,000,000
6. Landsat E-21272-21063 MSS Bands 6 and 7; SUN EL 09 AZ 048, acquired 17 July 1978, scale 1:1,000,000
7. Enlarged Landsat E-2192-21265 MSS Band 7; scale 1:450,000.

The image selected (Fig. 7.1; no. E 2282-21254 from October 31, 1975) contained high-quality digital data; that is, complete data in all four spectral bands and no data drop-outs, thus, no digital-data restoration was necessary. The enhancement procedures used on this image included contrast-stretching and spatial filtering. Image enhancement involves the modification of an image to alter its visual impact on the viewer. Its goal is to aid the analyst in the extraction and interpretation of pictorial information.

Contrast enhancement involves expanding the pixel values so as to maximise the range of grey values. To achieve this linear, histogram-equalized, histogram-normalised and space variant contrast stretches were performed at the Physics and Engineering Laboratory (PEL) Wellington (Fig. 7.2).

Edge enhancement was achieved by high-pass filtering in order to emphasize higher spatial frequences. This was done using a convolution operation which is useful for increasing contrast differences for the enhancement of lineaments.

Colour enhancement is applied to take advantage of the ability of the human visual system to discriminate a greater number of colours than it can shades of grey. In the display of multispectral images, coincident images in three different spectral bands are combined by displaying them in the primary additive colours of blue, green and red (Fig. 7.3). Subscenes of the lower Waipara area were examined on the

Figure 7.4

Landsat image of the North Canterbury-South Marlborough region. Color composite Landsat 2 image of 2192-21265, aquired August 2, 1975 PEL 13.

Figure 7.3

Close up enhanced color composite Landsat 2. Band 1,2, & 4. Stretch Band 24-127 H.E. of the lower Waipara area.

colour video monitor of the PEL image processing system and an enhanced colour composite of the whole studied area was reproduced as a colour photograph (Fig. 7.3).

Computer-aided classification mapping of the digital data was not employed in this study. Instead, it was felt that visual and stereoscopic interpretations permit the author more effectively to utilize his field experience and understanding of the mapping procedure; an interaction which is difficult to transfer to computer algorithm.

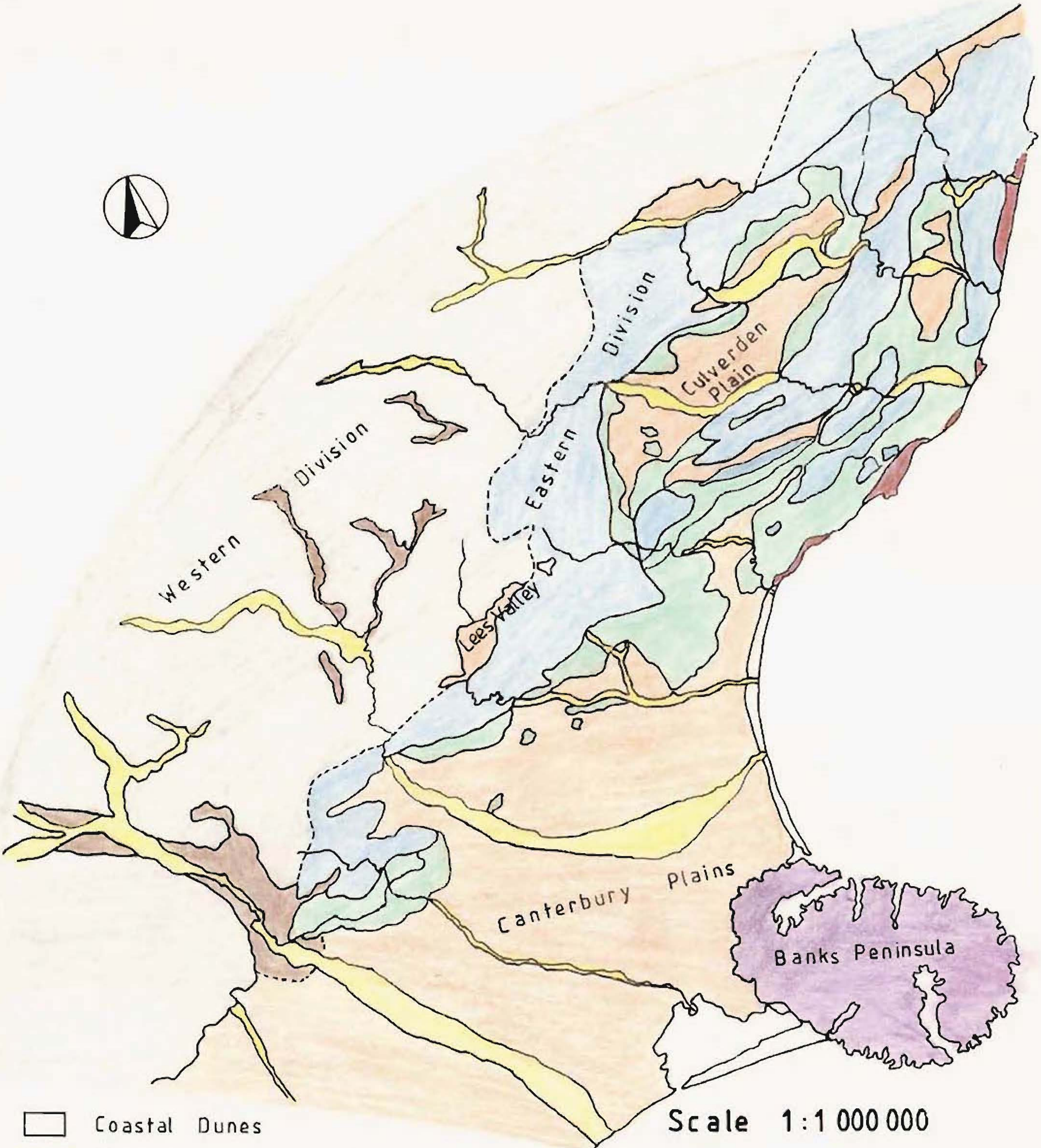
7.2 GEOMORPHIC LANDFORM CLASSIFICATIONS

Landforms in the north Canterbury - south Marlborough region fall into several classes. The fundamental image characteristics which were used for identification and analysis were tone, texture, pattern, shape, drainage and associated orientation corresponding to geomorphic classes. The author recognized nine broad genetic groups of landforms, which are represented in Plate 15.

In order to make later discussion clearer these units are briefly described before more detailed landform descriptions are given in the next sections. From east to west their characteristics are as follows:

7.2.1 RAISED MARINE TERRACES

The Late Pleistocene marine terraces border the eastern flanks of the coastal ranges rising from 10-275m above mean sea levels (Carr 1970, Ota et al. 1984). They are separated from coastal ranges by their gently sloping topography and distinctive tone and slope-break relationships. There is little or no coastal plain along the shore of



- Coastal Dunes
- River Terraces and Floodplains
- Wash Plain (fluvio / glacial)
- Raised Marine Terraces
- Quaternary Post-glacial Alluvium (footslope)
- Volcanic Cone
- Undulating Hill Zone in Cainozoic Covering Strata
- High Mountain Zone in Resistant Basement Rock
- Steeply dissected Mountain Zone in Resistant Basement Rock

PLATE 15 Map of geomorphic units of north Canterbury-south Marlborough, mapped from Landsat images.

the coastal ranges because the actively growing folds are so young that erosion has not reduced the terrain.

7.2.2 COASTAL DUNES

Dunes sand and coastal lagoonal deposits border the eastern margin of Canterbury plains extending as far as 5 Km inland. Coastal dunes were difficult to distinguish from raised marine terraces and with little topographic and textural contrast, these two categories have not been differentiated.

7.2.3 QUATERNARY POST-GLACIAL ALLUVIUM

The Canterbury and Culverden Plains surrounding the mountains consist of Quaternary glacial outwash gravel, kame terrace deposits, local fan and river aggradation gravels and morainic deposits known as the Burnham Formation, (Gregg 1964). The building of the two plains is related to the events of the Late Pleistocene and to the uplift of the Southern Alps in the Kaikoura Orogeny. These plains were dissected by rivers during the warmer climates of the interglacial, interstadial and post-glacial period when vegetation regenerated at higher altitudes in the Southern Alps. Remnants of moraines and outwash deposits of early glacial advances are present in the inner parts of the two plains.

7.2.4 VOLCANIC CONE

The volcanic edifice forming Banks Peninsula, is clearly indicated by the topographic prominence consisting of the remains of two large stratovolcanic centers, Lyttelton and Akaroa, both of Miocene age. Volcanism occurred more or less continuously over a period of 6.2 Ma, beginning approximately 12 Ma ago and ending about 5.8 Ma ago (Sewell 1985). During the glacial periods of the last 2 million years thick blankets of loess, up to 20 metres thick were deposited on the irregular surface of Banks Peninsula.

7.2.5 RIVER TERRACES AND FLOOD PLAINS

This unit is the area occupied by the river bed and surrounding flood plain and terraces. Most of the major rivers have broad gravel-bed channels. Consequently, they are not difficult to detect on the near-flat lowland. Note how the narrow, bed rock rivers in the eastern highlands become broad, braided rivers on the plains (Fig. 7.4).

7.2.6 WASH PLAIN

The major valleys in the Southern Alps landscape shows typical landforms of glacial origin. The fine fluvial dissection on the Southern Alps sharply contrasts with the coarse dissection patterns of the uplands and the broad flat lowlands formed by glacial processes. Fluvio/glacial processes since the Pleistocene glaciation have formed the very broad wash plains occupying the lowland.

7.2.7 UNDULATING HILL ZONE IN CENOZOIC COVERING STRATA

The Cenozoic covering strata are restricted here to that southwest trending belt of en echelon folds and faults within the sedimentary section of the north Canterbury -south Marlborough basin. The various patterns of dissection in the undulating hill zone correspond closely with differential resistance to erosion in the covering strata.

7.2.8 STEEPLY DISSECTED MOUNTAIN ZONE IN RESISTANT BASEMENT ROCK

The resistant basement rock (Torlesse Supergroup) can be divided into two groups by morphology and location. That is, the more steeply dissected western division is notable for coarse-textured dissection and fault controlled drainage pattern. Two kinds of landscape features, high mountains and deep valleys, dominate the high ranges of the Southern Alps. They result from the interaction of two powerful processes working in opposite directions the slow uplifting of great segments of land in response to enormous pressures in the earth's crust, and the carving of the valleys by erosion (particularly glacial erosion) into the land as it rises: These processes have affected and continue to affect this western division.

The geomorphic character of the eastern division has a distinctly different erosion pattern, lineament and stream density to areas of Torlesse Supergroup in the western division of apparently similar lithology. The eastern division consists of high mountains and closely-spaced valleys which form a fine-textured dendritic pattern (Fig. 7.4).

The young uplift of the Southern Alps has resulted in a rather simple drainage system with rivers draining northwest and southeast away from the divide. The pronounced lithological contrast between east and west slopes of the Southern Alps controls the nature of the rivers and the depositional landforms (Adams 1980a).

7.3 DRAINAGE PATTERNS AND THEIR CHARACTERISTICS

Interpretation of Landsat images for terrain applications relies upon drainage pattern/texture identifications and analysis. It is

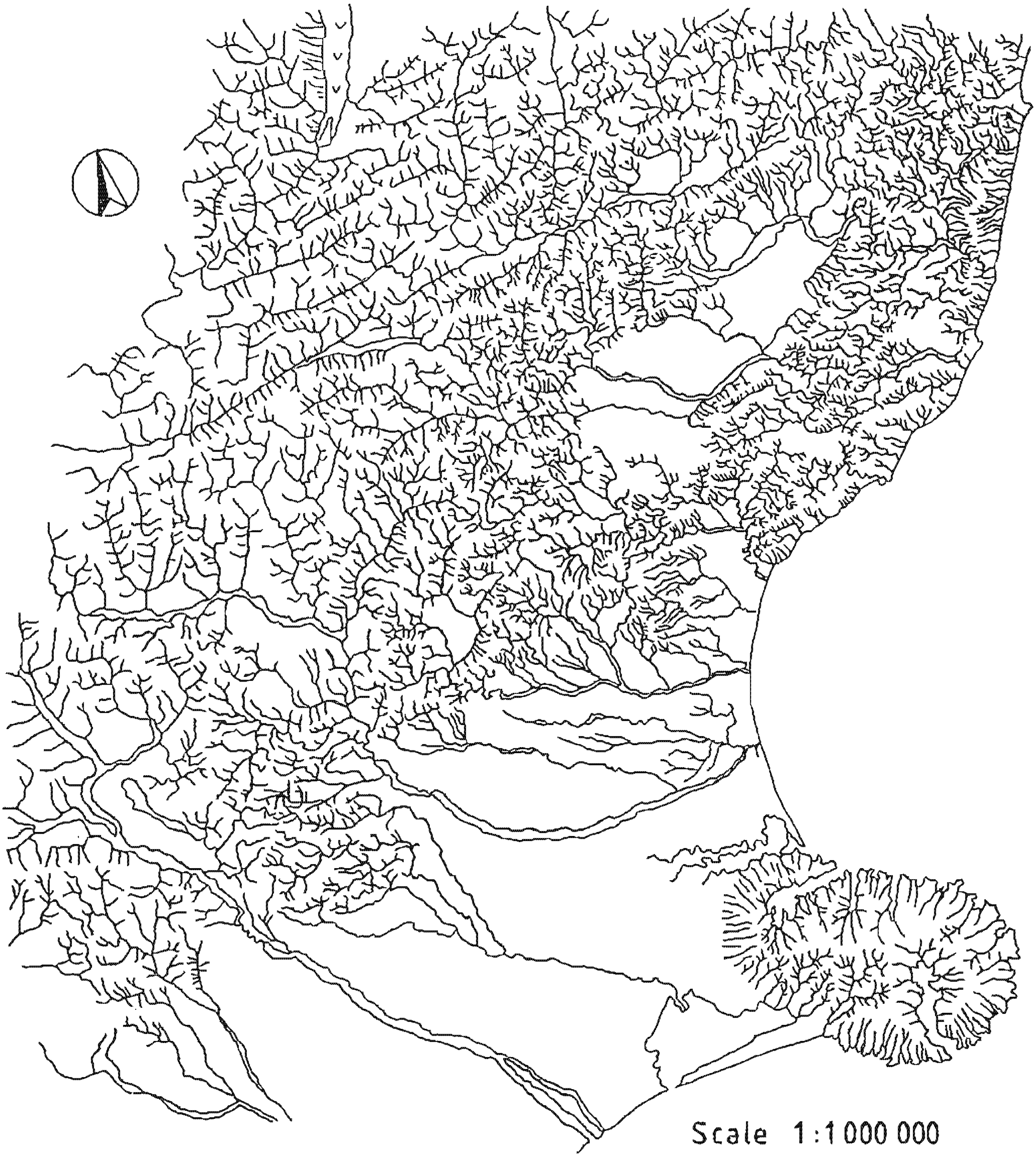


Figure 7-5

Drainage networks of north Canterbury - south Marlborough. Network was compiled from streams shown on Landsat images.

obvious from drainage networks (Fig. 7.5) that significant correlations exist between the physical properties of rock and the relief influenced by landscape evolution portrayed by drainage pattern, density and topography. Additionally, it is often possible through detailed analysis of the drainage networks (Fig. 7.5), including local variations in the stream pattern, to establish definitive criteria associated with active tectonic folding and faulting. These unstable areas are ordinarily accented by the presence of anomalous patterns formed by fluvial erosional processes. The well known alignment of drainage along fault traces, is only one of the diagnostic patterns and landscape features among the criteria used. The full range of criteria associated with active folding and faulting is considered to be one of the major findings in this study, using Landsat images with a scale of 1:1,000,000.

Six to nine different drainage patterns can be detected. These include a radial drainage pattern on the volcanic cone where many of the distinctive valley systems on present-day Banks Peninsula probably formed by persistent erosion in the head waters of a major radial drainage pattern. Structurally controlled fine and coarse dendritic patterns are recognized inland, composed of an initial curved trellis and a sub-parallel pattern. A directional trellis pattern following the fault traces is obvious in the upper left corner of the scene. In general, stream junctions in this drainage pattern are nearly at right angles. Parallel to sub-parallel patterns are adjusted to regional dip directions. The trunk stream flowing approximately normal to the regional dip direction has dominated the coastal zone. Braided river pattern developed on the broad Canterbury and Culverden Plains and inland glacial valleys. Finally, a local meandering pattern in a low and high relief terrain is distinguished.

7.4 GEOMORPHIC LINEAMENTS

Landsat MSS imagery proved particularly useful for revealing lineaments within the study area (Oliver 1978). Some of these structures were known or suspected from ground studies, but the majority were revealed for the first time during the Landsat investigations (Fig. 7.6). While classifying the lineaments, the author has taken into consideration both physical and genetic parameters. On the basis of physical appearance, lineaments have been divided into:

I Curvilinears - which could represent various fold closures and/or erosional/drainage characters;

II Linears - which could represent fracture zone and formational boundaries. Some linear features which do not fall into the above category, yet are well defined on the image, perhaps represent new local faults.

These lineaments have further been classified as major, intermediate and minor depending on their length of 30 km or more, 30 km to 5 km and less than 5 km respectively.

The linear and curvilinear lineaments shown on (Fig. 7.6) can be interpreted using relatively standard photogeologic or geomorphic analyses such as has been done for many years in interpreting data from aerial photographs. Most of the lineaments observed are undoubtedly related to the horizontal strain field that is a result of New Zealand's position on the boundary of two major crustal plates, the Pacific Plate and the Indo-Australian Plate. Oceanic crust is being subducted in the Hikurangi Trough whose southern end is off the

Figure 7-6

Structural - lineament map of north Canterbury - south Marlborough area interpreted from Landsat images.

coast of Kaikoura Peninsula. Continental crust is present in the adjacent plates along the transform Alpine Fault, which splays into four major fault zones in the northeastern side of the study area (Fig. 7.6). Since the Oligocene, this plate boundary has been subjected to a complex pattern of regional compression, extension and transcurrent displacements, collectively referred to as the Kaikoura Orogeny.

Lineaments observed in the image are almost entirely confined to the basement and Cenozoic covering strata. Many of the major lineaments are known faults predominantly oriented in an east-west to northeast-southwest directions. The second set of lineaments striking in north-south to northwest-southeast directions, is visible in the high ranges of the Southern Alps which correspond to mapped faults and glacial valleys. Most of the minor lineaments are predominantly parallel to the major and intermediate lineaments, are constrasts either between lithologic boundaries or small faults. Subsequently, according to the distribution of various lineament types, the region may be divided into four tectono-physiographic zones (Fig. 7.7). Their characteristics are described as follows:

7.4.1 WESTERN ZONE

The prominent lineaments were mainly those controlled by major straight valleys (Fig. 7.6). East and northeast trending fault traces marked by drainage alignments are responsible for the offset in stream and surface exposures in the upper left corner of the scene. The Hope, Clarence, Awatere and Elliot active dextral faults splaying northeast from the main Alpine Fault are most readily observed on the colour composite images (Fig. 7.4). Walcott has shown from geodetic

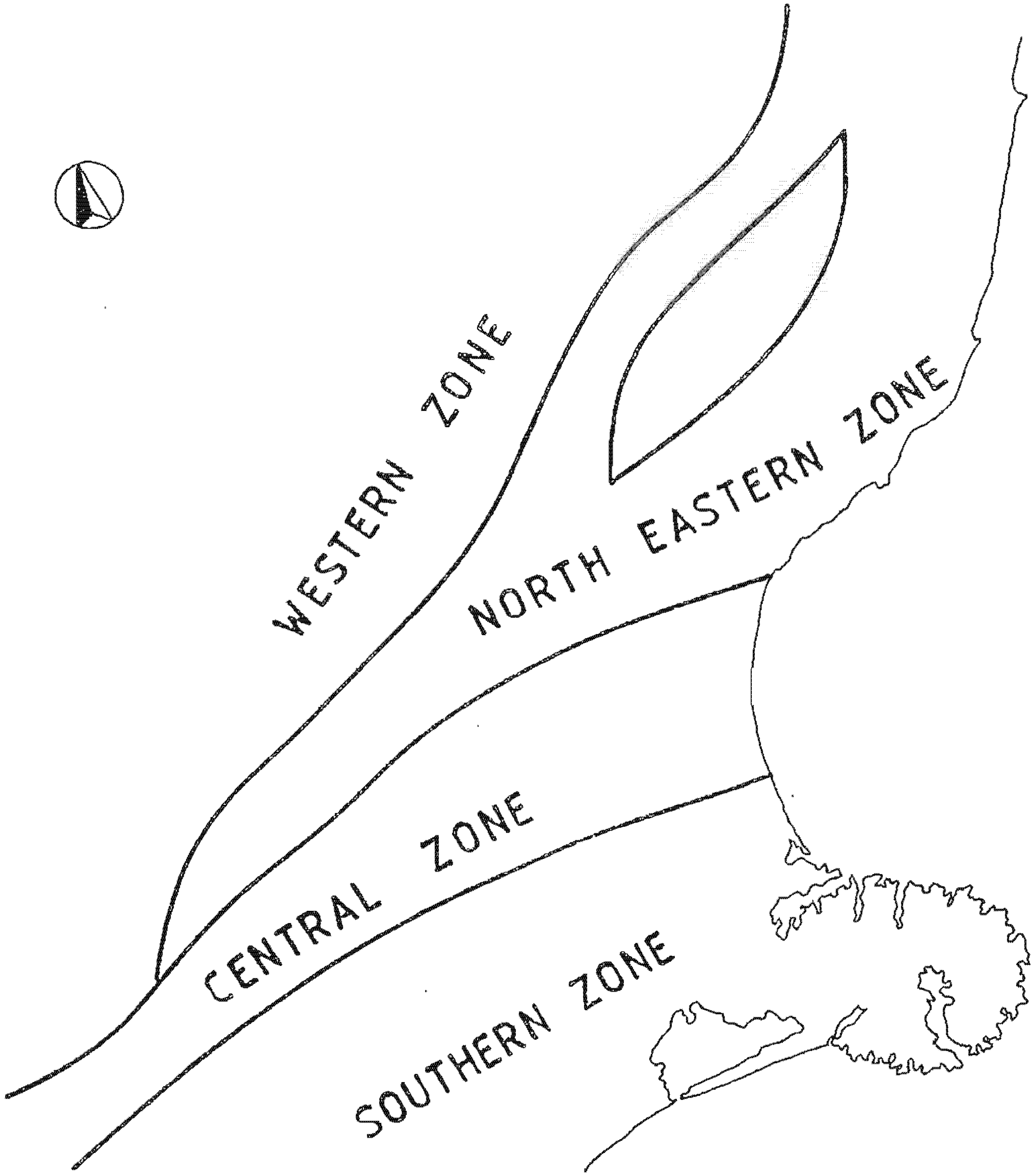


Figure 7 -7
Tectono - phsiographic zones of north Canterbury-
south Marlborough region

data that the major faults of the Marlborough shear zone only accommodate about half of the observed modern strain, and that "at least half the movement ..... may be made up of displacements on numerous minor faults, rotation of blocks between the major faults or by ductile flow in the crust" (Walcott 1978, p.153). A second set of lineaments striking north is recognisable in this area. The largest of these lineaments cuts across the other major lineaments. Examination of(Fig. 7.6) shows surprisingly few examples where lineament sets appear to have been right-laterally displaced along other lineaments.

Major lineaments in the lower left corner of the scene correspond to the known faults called the Harper and Torlesse Fault Zone, with a north-east strike. From Landsat images three lineament directions occur in these mountains. The dominant lineament trend is parallel to the known faults. The second set of lineaments strike approximately north-south, and the third, which are obvious in this area, strike north-west, represent glacial valleys, probably along fault crush zones.

7.4.2 NORTH-EASTERN ZONE

In the northeastern coastal belt of the region there are two prominent drainage patterns and lineament trends with similar orientations to those observed in the eastern division of the Southern Alps (Fig. 7.6). This distinct geomorphic feature indicates that these two areas may be of the same tectonic regime. Figure 7.6 shows an obvious three lineament trend in this zone. The dominant minor lineament trend is sub-parallel to the known major fault (called Kaiwara, Hundalee and Hamilton Faults), and strike of beds in this area (Gregg 1964, Bradshaw 1972). The second set of lineaments strike approximately north-south, and the third, which are the least dominant, strike northwest.

7.4.3 CENTRAL ZONE

In the Canterbury Plains, north of the Waimakariri River, there are strong tonal-drainage lineaments, probably marking the surface trace of new fault traces, indicated by dotted lines (Fig. 7.6). No conclusive field evidence is known for a set of faults, but these lineaments follow onshore fault traces southwest of the offshore Motunau Fault recognised by Carter and Carter (1982). These set of subparallel lineaments strike east-west to northeast-southwest and appear to be responsible for the channel pattern changes in this area (Fig. 7.8). The discovery of these lineaments, and their identification as recently active features are of considerable significance in this study. Moreover, these lineaments are marked by a set of northeasterly trending concentration of earthquake epicenters just northwest of Christchurch city (Fig. 7.9). In fact, these lineaments mark the southern boundary of the Marlborough shear zone. Similar trending lineaments at the mouth of Rakaia and Selwyn Gorges, together with the recently active Porters Pass dextral fault zone encompasses the western margin of the central active zone.

7.4.4 SOUTHERN ZONE

Most of the lineaments in this zone appear to be man-made linears, such as roads or railroads, agricultural patterns or other nongeologic features. However, there is a prominent change in lineament concentration and seismic activity from south of the

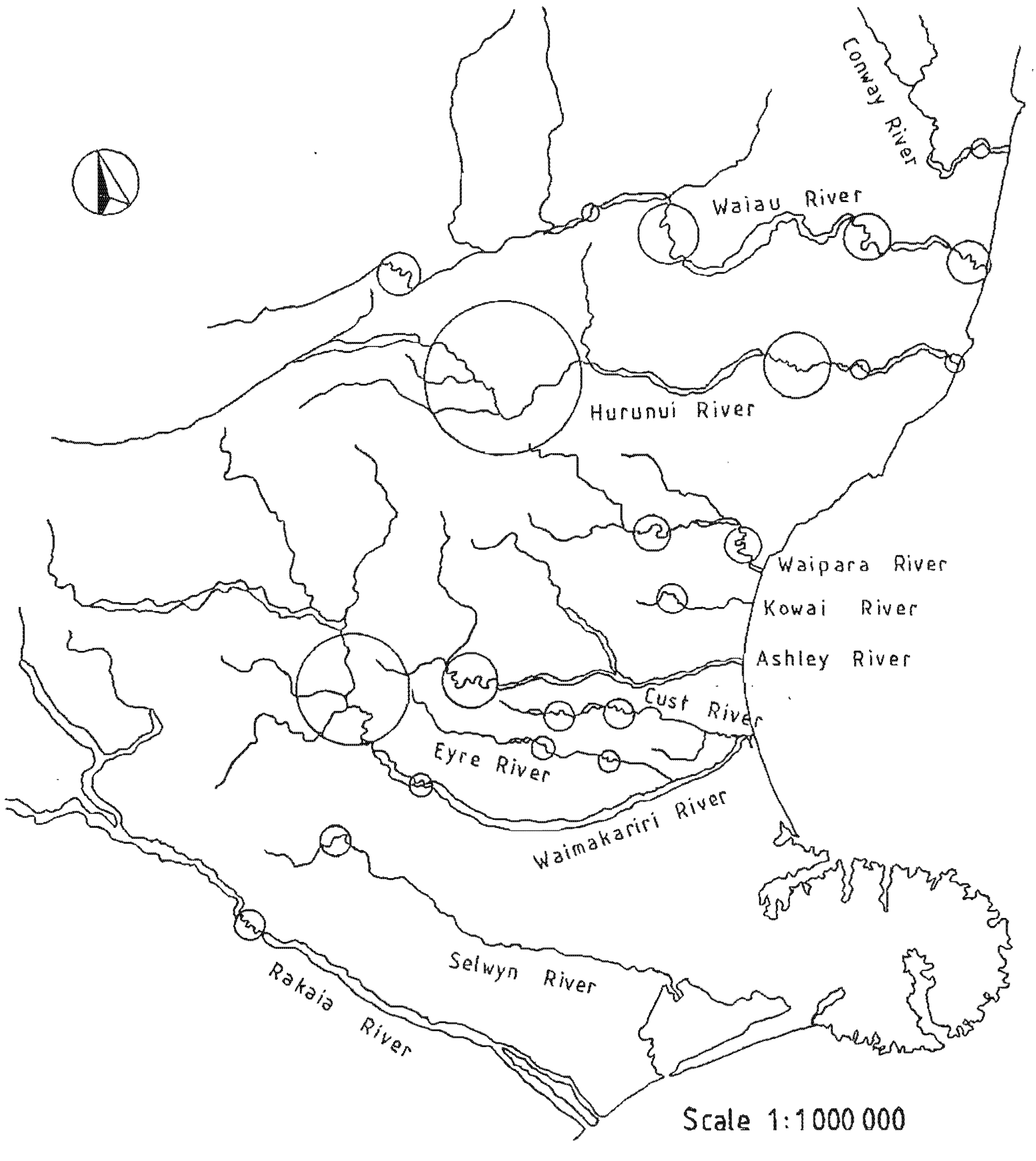


Figure 7 - 8

Anomalous stream pattern in very low and high relief terrains, observed from Landsat images of north Canterbury - south Marlborough. Circle indicate position of stream pattern change.

Waimakariri River. Within this zone seismic activity consists of scattered shallow, low intensity earthquakes (Fig. 7.9).

7.5 SEISMIC ACTIVITY

Use of small scale Landsat MSS imagery and crustal seismicity map solutions for regional tectonic interpretation of this area was described by Carter and Carter (1982). These workers demonstrated that seismic activity shows a regional change in pattern across the Motunau Fault. Within the zone between Motunau and Hope Faults, high intensity and deep earthquakes are largely restricted to an area in the northeast, the southern limit of which corresponds to the termination of the Motunau Fault against the northerly trending faults at the head of the Hikurangi Trough. South of the Motunau Fault, seismic activity consists of scattered shallow, low intensity earthquakes. Two active dextral fault traces known as the Ashley and Pegasus Bay Faults were noted by Lensen (1977) and Carter and Carter (1982), south of and subparallel to the main Motunau Fault. But fault traces were not recognised on the Canterbury Plains.

The present study developed that work a stage further by combining photo-interpretation of a 1:1,000,000 scale MSS band 7 image with enhanced images to examine critical areas in detail. One of the most striking results of this analysis has been the discovery of numerous, previously unmapped regional linear features in areas considered to be reasonably well mapped (Fig. 7.6). Some of these lineaments are marked by a conspicuous northeasterly trending concentration of earthquake epicentres just northwest of Christchurch city (Fig. 7.9). While the author was writing this chapter an earthquake measuring 5.25 on the Richter scale occurred on the 9th of

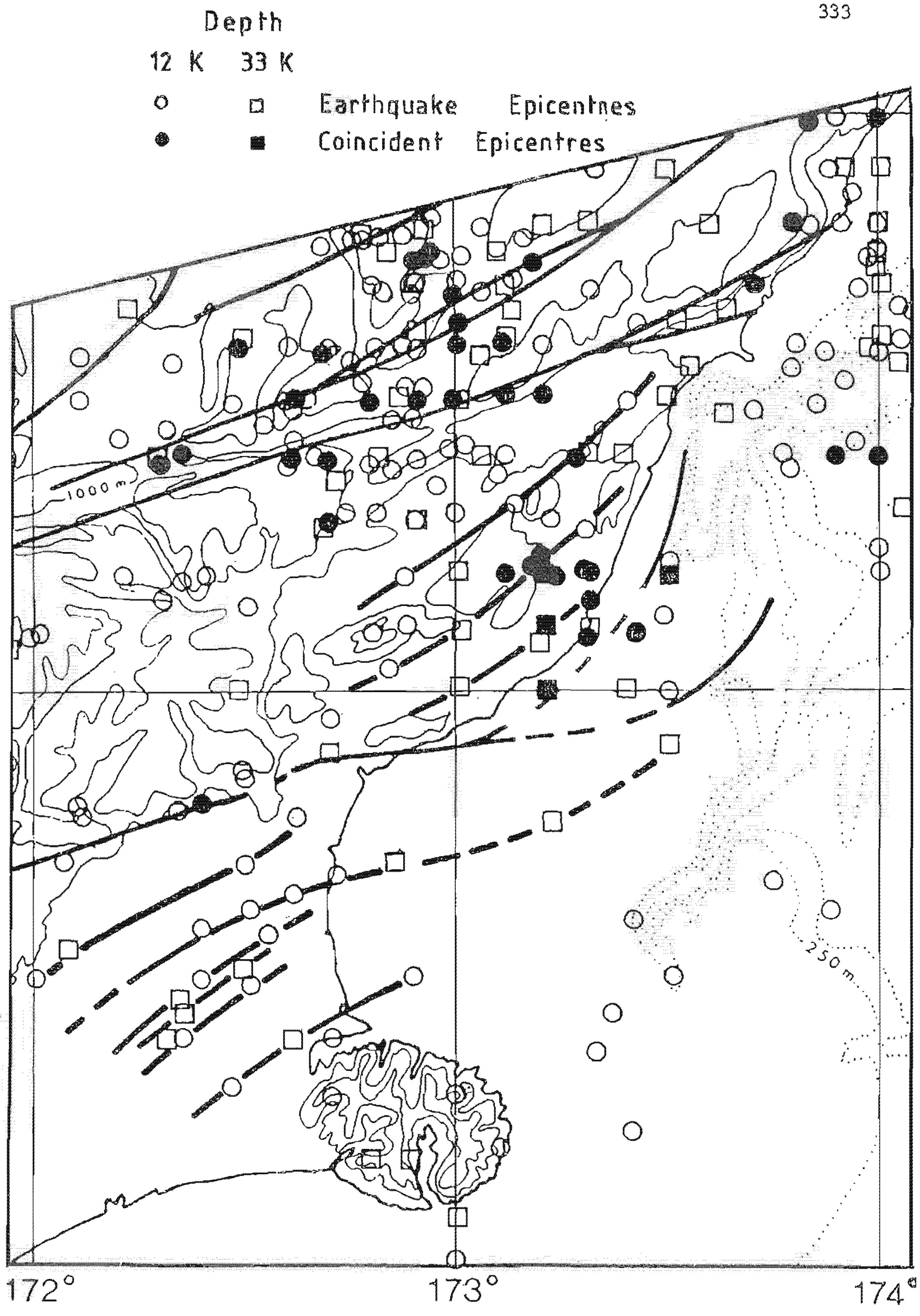


Figure 7.9

Crustal seismicity in the North Canterbury - South Marlborough region, 1942-1979 inclusive (data supplied by Geophysics Division, New Zealand D.S.I.R).

March 1987 and was centered in Pegasus Bay about 80 km northeast of Christchurch, an area currently assigned a grade 2 earthquake risk rating (Clark et al. 1965). This area could be struck by another earthquake any time similar to the moderate one which jolted the region. Thus, the above grading may require reassessment.

7.6 THE ANTECEDENCE CONCEPT

The origin of the anomalous river pattern shown in (Fig. 7.8), can be interpreted in terms of an hypothesis of antecedence, as can many of the smaller streams in north Canterbury - south Marlborough region. Since the pioneering work of Cotton (1913), there has been considerable debate among New Zealand geologists and geomorphologists as to the nature, origin and structural relationships of a number of New Zealand rivers (notably Hutton 1905, Cotton 1913, Speight 1912 & 1915, Adkin 1930, Jobberns 1937a, 1937b, Gregg 1950, Powers 1962, Fair 1968, Carson 1984, Campbell & Yousif 1985). The structural barriers crossed by New Zealand rivers are of different ages, and most of them involve late Tertiary as well as late Quaternary movements. The anomalous meandering valleys through the barriers are of different ages and are the result of different kinds of processes.

The geologic history of the major rivers begin with the emergence of the Southern Alps from the sea in the Kaikoura Orogeny. The main stem of the major rivers began with a consequent course that drained eastward on the newly formed structures. That original consequent course was different from the present one. During the early Kaikoura Orogeny, the east flowing consequent rivers became interrupted by block faulting that formed basins and ranges across the river courses. At most of the structural barriers, the ancestral course may have been started when the impounded river overflowed low places on the rim, and has been able to maintain its course by eroding downward as the uplift progresses and cutting the deep rugged gorge into the mountain ranges. East-draining rivers like the Rakaia, Waimakariri, Ashley, Waipara, Hurunui and Waiau occupied a gravel-filled valley widened during the last glaciation. The present mountain ranges marked by abundant earthquake epicenters, recent fault scarps, and measurable present-day tilt indicate that the ranges continued to rise through late Quaternary time to the present day.

River patterns were studied from Landsat images for the purpose of obtaining information on the morphology of channels. Similar to the method used for lineament classification, the rivers are divided into morphologically homogeneous reaches (free braided, entrenched meander, free meander and fault-controlled rivers). These rivers have further been classified as major, intermediate and small depending on their length and discharge. Rivers within active ground structure are deeply entrenched in the sedimentary cover and have meandering like valley with a narrow floodplain (Campbell & Yousif 1985). Circles in (Fig. 7.8) indicate the positions of anomalous meandering patterns which may be related to the nature and relative activity of currently active ground deformation. The folding, faulting and tilting have continued as repeated, even though minor movements, while the river has been cutting its meandering valley, and probably are still going on.

7.6.1 MAJOR RIVERS

The Hurunui and Waiau, together with the Rakaia and Waimakariri, are the major rivers of the north Canterbury -south Marlborough region, with their sources high up in the Southern Alps (Fig. 7.8). The broad Canterbury and Culverden plains and the smaller Lees Valley are renowned for their large braided gravel-bed rivers. All these rivers display a transitional pattern, often exhibiting a distinct downstream progression from a braided to a meandering appearance.

The Hurunui and Waiau are the most complex of all those studied, because they experience many changes in pattern along their courses apart from the obvious effects of Hope Fault controls on river alignments (Fig. 7.8).

This investigation showed that most of the meanders on the Hurunui and Waiau Rivers are confined to the basement block uplifts (Plate 15). The crest of these mountain blocks seem to indicate the presence of an air gap, a relief of initial topography that would guide the consequent drainage during the early stage of the deformation that ultimately raised the range and enclosed the basin. In the later, more intense stage of this deformation the shape of the surface changed very considerably and entrenched meandering valleys became established. In my opinion, this phenomenon is attributable to the proximity of the deformation zone, and one should expect the largest number of meanders within growing tectonic structures.

The transformation of the Hurunui and Waiau to a fully braided rivers at the start of the broad aggraded depression in which Culverden stands is shown in (Fig. 7.8). Exposed broad straths that are bevelled across valley floors during extended periods of equilibrium riverflow together with thick overlying aggradation gravels are recognised only along the aggraded depressions. Continuing incision in the narrow, rugged gorges is consistent with the uplift rates of that part of the mountain ranges, in contrast to the slow relative subsidence in the basin. Whereas the correlation of the meander anomalies with actively rising blocks is very obvious in the examples just discussed, similar but more subtle indications of deformation may be observed on the floodplain surfaces themselves.

Below the mountain front the Waimakariri River changes in character showing drainage anomalies in a very low relief terrain, as is by no means unusual on major rivers draining east from the Southern Alps. Immediately downriver from its upper gorge (Fig. 7.8), the Waimakariri River displays a well-defined braided pattern prior to the lower meandering anomaly. This second meandering course is confined into a deep and narrow gorge within the old valley, cutting through a thick mass of strongly indurated greywacke type rocks of Torlesse Supergroup. The present abrupt transformation to meandering in this reach is not hard to understand. A strong tonal lineament, probably marking the surface trace of an active fault which uplifted the basement strata in this area has been mapped from Landsat images (Fig. 7.6). The lower Waimakariri Gorge shows many similarities to the Rakaia River which has been discussed in the previous chapter. This segmental variation in channel pattern and width may be related to lithologic and/or structural uplift. Both the Rakaia and Waimakariri continue in their braided habit to the coast as they emerge from their gorges (Fig. 7.8). The Waimakariri, however, changes character abruptly 5 km from the coast and becomes a single channel stream in contrast to the Rakaia, which is incised below the Canterbury Plains near the river mouth, forming two channels (Fig. 7.8).

Field checking has revealed a marked development of several very young terraces, abandoned river channels and valley bend which can be traced through the northwest of Christchurch city, where the

Waimakariri River swings from a south-easterly to an easterly course.

7.6.2 INTERMEDIATE RIVERS

The Waipara and Ashley are the intermediate rivers, with their sources in the eastern foothills of the Southern Alps.

From the headwaters to the sea, the Waipara River crosses two structural barriers in antecedent gorges and the bed form alternates between braided reaches of aggraded flood plain and deeply incised channels in bedrock, which have been discussed in the previous chapter.

The Lees Valley extended along the trend of the Southern Alps and is formed by a structural basin floored by glacial outwash and river aggradation gravels laid down during the last Pleistocene (Otira) Glaciation and is about 60 km^2 in area (Plate 15). On either side of the valley exposed straths have developed, demarcating the boundary between the Torlesse Supergroup and the overlying aggradation gravels (Fig. 7.10). About 16 m of tectonically induced downcutting has occurred between the exposed strath and the major strath forming under the present floodplain of the Ashley River (Fig. 7.10). The Ashley River flows south to south-west across the Lees Valley and is characterized by a wide gravel-bed braided pattern. An angular unconformity can be seen between the two units; this has been brought about by renewed movement along the fault bounding eastern division of the Southern Alps in the interval after the deposition of upper unit.

The examined part of the Ashley River in the eastern division of the Southern Alps (Fig. 7.11) is incised up to about 350 m into a mass of strongly indurated, mostly graded-bedded greywacke and argillite of Torlesse Supergroup. The continuous uplift of the eastern foothills of

Figure 7.10

Exposure of strath (arrowed) and overlying aggradation gravels through the Lees Valley upstream from the Ashley Gorge.

Figure 7.11

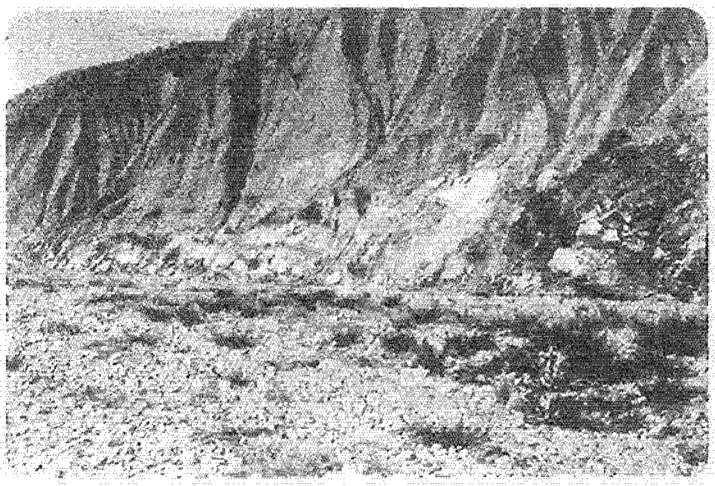
The gorge of the Ashley River, cut through the uplifted Torlesse Supergroup of the Southern Alps. Note the entrenched meanders.

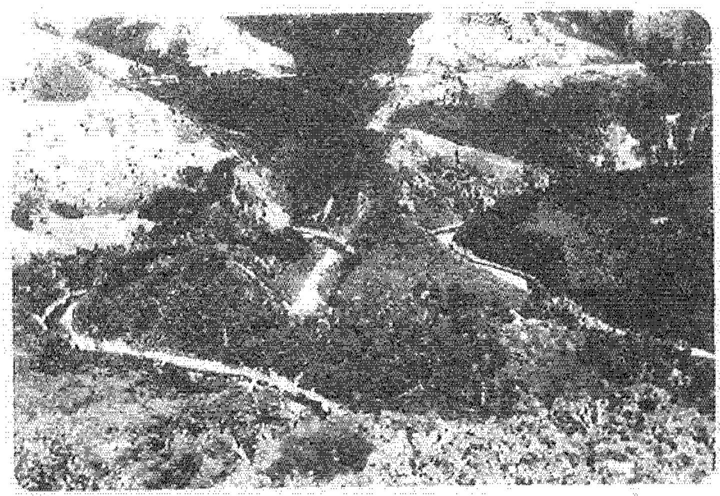
Figure 7.12

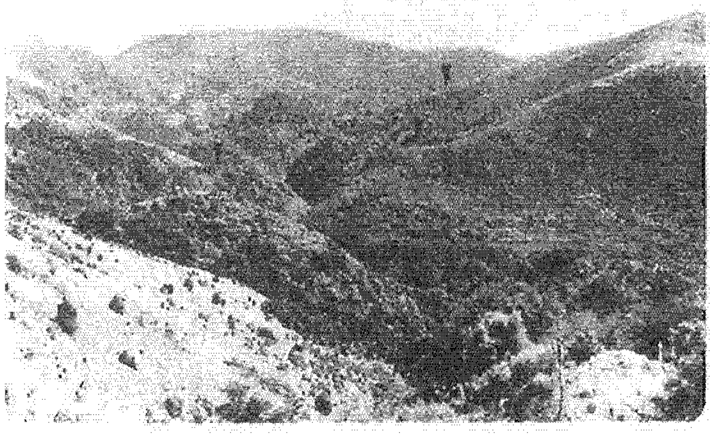
The Ashley River has constantly cut down through the rising of the Southern Alps and, coupled with regime changes associated with climatic changes, has produced a spectacular series of terraces (arrows) which are shown in this photo.

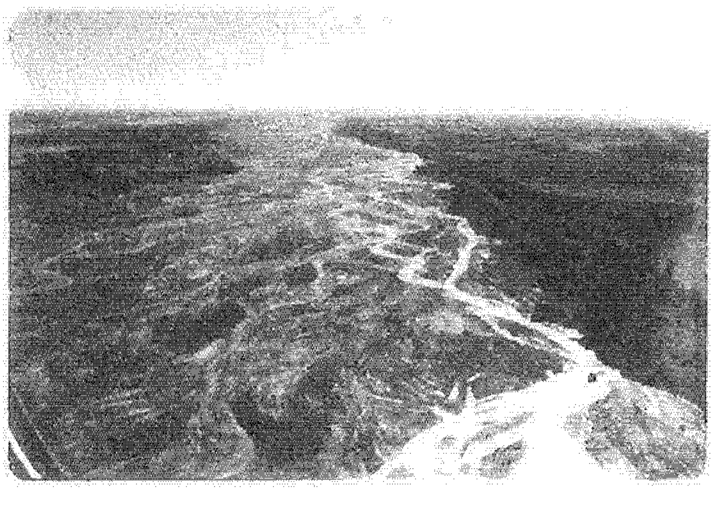
Figure 7.13

Typical braided river of the Ashley River..









the Southern Alps is indicated by the presence of a sequence of flights, erosional terraces, entrenched meanders and contrasts in channel pattern apparently resulted from uplift of the range (Fig. 7.12). Downstream, the valley opens out into a second wide floored gravel-bed reach and flows east across Canterbury Plains (Fig. 7.13).

7.6.3 SMALL RIVERS

The Kowai, Cust and Eyre Rivers are much smaller, rising in the foothills, and dissecting in their lower courses various sequences of Quaternary alluvium. Although small, they have high values of specific stream power. All the examples discussed have been powerful rivers that readily erode soft and moderately resistant materials and act as connecting links between tectonically active and inactive parts of the fluvial systems. Here attention is turned to some of the smaller rivers observed from Landsat images showing segmental variations in channel pattern (Fig. 7.8). These variations may be related to structural controls.

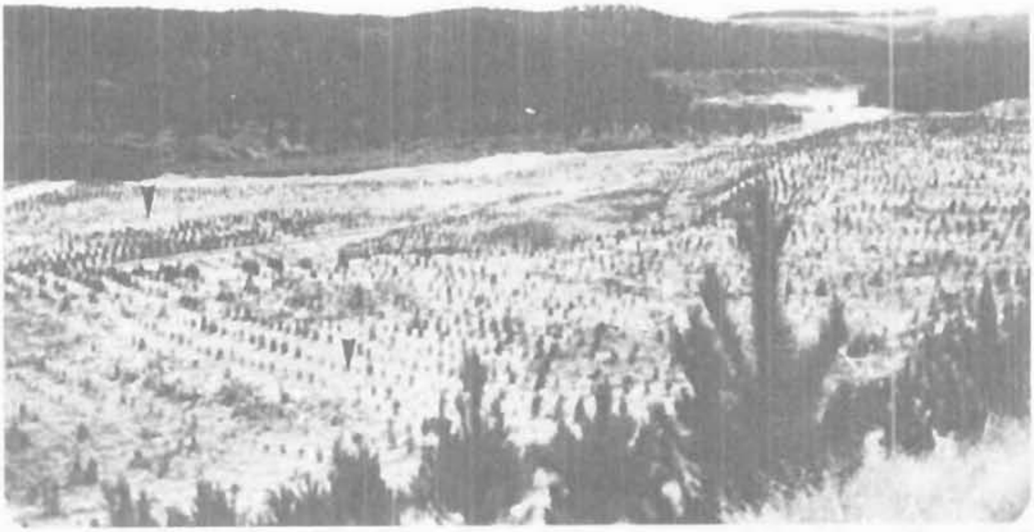
The Kowai River (south branch) shows many similarities to the case studies of Kate and Carrington Creeks which have been discussed in previous chapters. A survey of its course showed the following pattern. Upstream, a consequent reach flows longitudinally along a synclinal trough then gives way to a relatively wide stretch of valley floor (Fig. 7.14), which broadens by lateral shifting. Downstream, the valley becomes more confined by an incised inner meander reach (Fig. 7.15) cutting across an anticlinal axis before opening out into a second wide floored reach with a much wider bare channel bed leading to the confluence with the Kowai River (north branch). This valley-in-valley form persists all the way down to the meandering

Figure 7.14

Synclinal reach of Kowai Creek showing a wide stretch of valley floor (arrows).

Figure 7.15

Landforms along the meandering reach of the Kowai River. Note the incised inner meander and flights of erosional terraces (arrows).



anomaly, although less well marked in certain reaches.

The river was probably diverted to a new course where the synclinal structure closes round the southwesterly plunging axis (Plate 17). This means the south branch was running parallel to the synclinal trough shown by the air gap and was once joined to the north branch. This new course of the river (meandering reach) is consequent on the new rising anticline. The contrast between the two types of reach, and the transition from the broadly rounded older valley profile to the deeply incised inner meander was apparent in Figures 7.14 & 7.15 and may be related to changing relationships of stream power to uplift rate.

The Eyre River is almost identical to the Cust River, where both rise in the foothills with similar discharge and incised within the same lithology in a very low relief terrain (Fig. 7.8). Two well-defined meander traces, judging by the sharply curved bend and narrower floodplain were apparent on each river course, observed from Landsat images and the topographical maps. Upstream of these meandering patterns they displayed a definite braided appearance. It is sufficient here to indicate that this contrast between the two parts of the channel pattern may be related to an east-northeasterly trending set of probable fault traces (Fig. 7.6), and perhaps the faulting has resulted in river ponding where cutoffs, accompanied by channel widening have been far more common in this area. Such processes have been shown by Gardner (1973, 1975) to produce similar patterns in experimental sand flumes.

7.7 MAPPING FROM LANDSAT IMAGES

The study of landforms of regional or tectonic significance, using topography between formational boundaries that separate rocks of different degradation rates together with lineament orientation and density upon which are developed drainage systems of contrasting patterns, are among the criteria used for interpreting geomorphic and geologic features from Landsat images (Plate 16 & 17). Several criteria which may be employed to determine the direction of dip of beds (as determined by the Rule of V's pattern), to distinguish anticlinal folds from synclinal folds, and to determine the direction of plunge of folds. Plunge of the folds is mapped by tracing beds that wrap around the plunging ends of the structure. Fault-bounded mountain ranges are characterized by triangular facets.

An area of approximately 5,062 km^2 from Ashley River in the south to the Hurunui River in the north was used for a target study to develop available methodology for the rapid production of small-scale geological and geomorphological maps from Landsat imagery (Plate 16 & 17).

The target area includes, but extends beyond, the lower Waipara study area, allowing for the extrapolation of methods and observations established within the field checked area to be applied to a related geological environment interpreted from imagery only.

Visual comparisons between the enlarged Landsat imagery (E-2192-21265) MSS band 7 (near infrared) and RBV imagery (Appendix 4) were made in order to evaluate the role of Landsat images in geologic mapping. Because of the high relief on the coastal range, and the complexity of the structure, high resolution is important, therefore the RBV system is the most suitable. The Landsat RBV system scene is

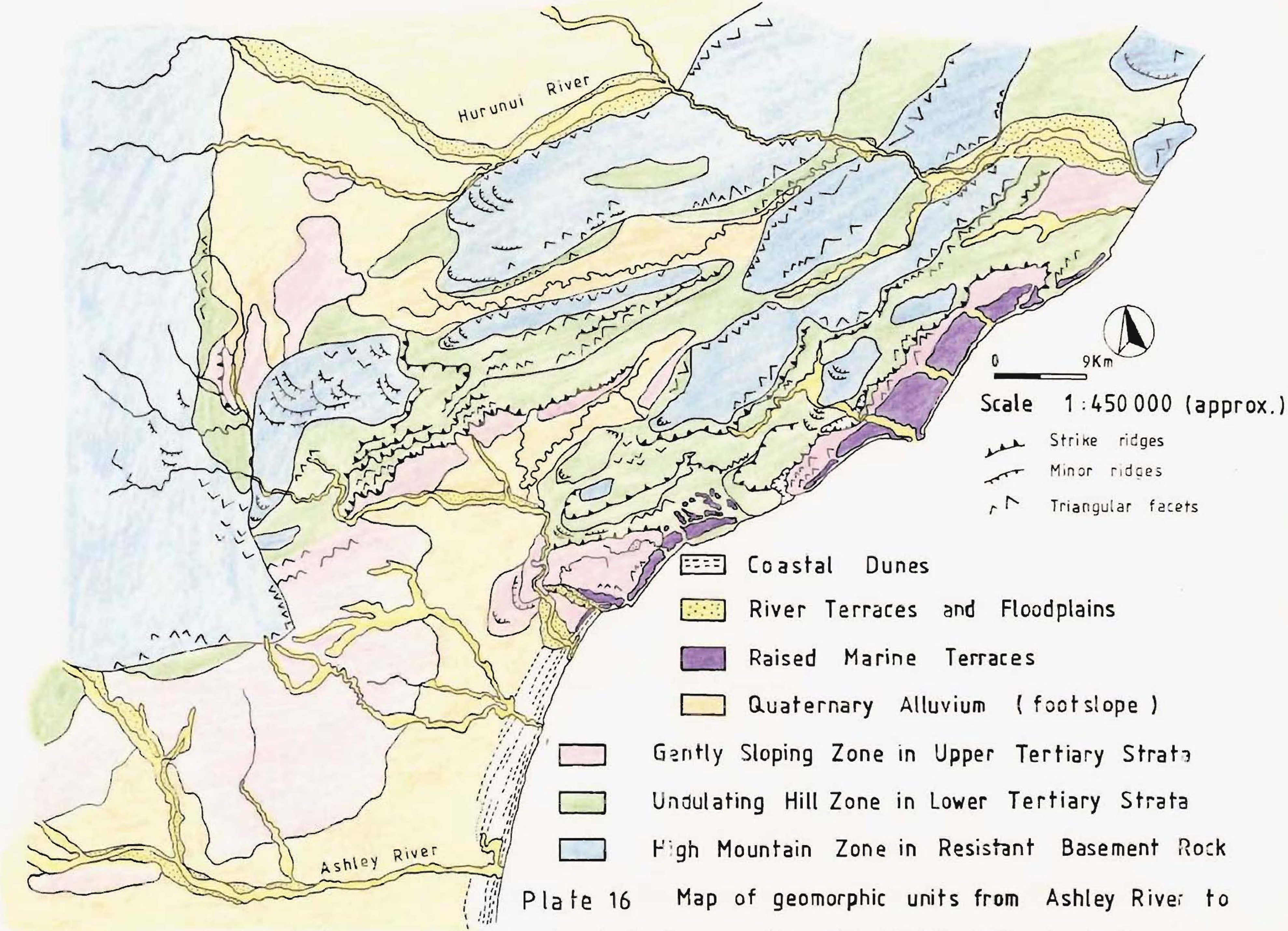


Plate 16 Map of geomorphic units from Ashley River to Hurunui River, mapped from enlargement Landsat imagery (E-2192-21265) MSS band 7.

in the visible and near infrared region of the electromagnetic spectrum (0.5 to 0.75 microns). The spatial resolution of the RBV is 38 meters, approximately twice that of the MSS system (Appendix 4). However, photogeological features such as drainage systems, bedding, resistance, texture and tone are better defined and therefore result in more accurate geological interpretations than normal size MSS imagery.

Most of the geological and geomorphological categories recognized from the enlarged MSS image (Appendix 4) can also be seen, though less obviously, on the RBV image. This can be seen by comparing these two images. An optimum way of utilizing both systems is to use the enlarged MSS imagery for drawing the actual maps and to include additional information by comparing this image with the RBV image. It was found that the geologic map produced from enlargement MSS imagery of an approximate scale of 1:450,000 was broadly similar to the published maps mapped by Gregg (1964) and Wilson (1963).

7.8 MAPPING THE COASTAL RANGE OF THE WAIPARA REGION

Comparison of the geomorphic map of the target area with the geologic interpretation map shows that topography is controlled by geologic structure. The canoe like ridges and valleys indicate that the bedrock has been folded and consists of stratified sedimentary rocks with different degradation rates. Mappable rock units are readily designated on the Landsat images but the rock types cannot be identified without field checking, or referring to geologic maps and reports. The southwest trending mountains and valleys are formed by anticlines and synclines, respectively. The eroded anticlines are

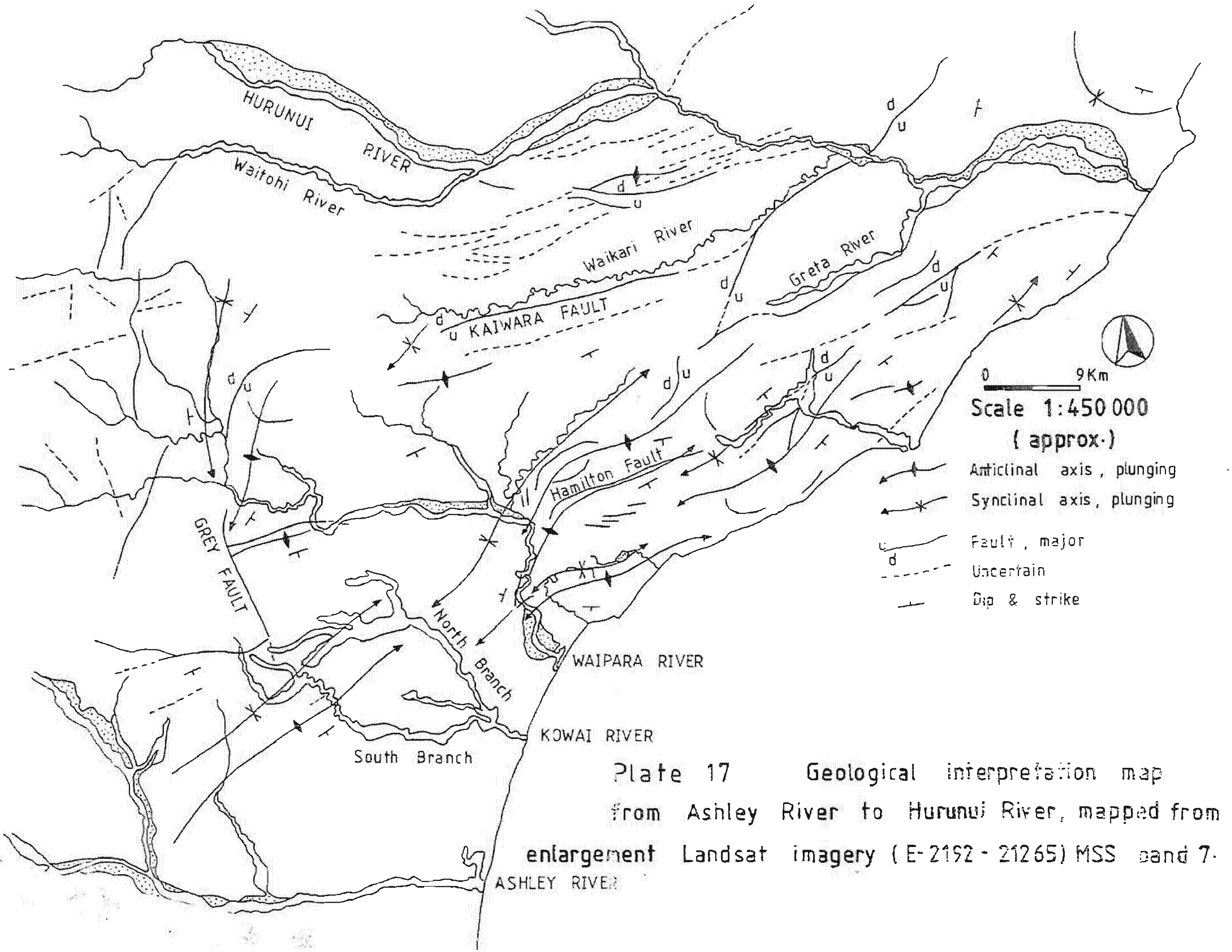


Plate 17 Geological interpretation map from Ashley River to Hurunui River, mapped from enlargement Landsat imagery (E-2192 - 21265) MSS band 7.

flanked by a well-stratified resistant unit that forms distinctive triangular facets and long strike ridges. Most fault-bounded mountain ranges indicate rejuvenation, as sharp linear contacts between the Quaternary alluvium and the mountain front are evident on Landsat images.

Plate 17 shows both large and small faults mapped by Gregg (1964) and Wilson (1963). The Hamilton Fault was known from ground studies prior to the Landsat investigations although its full extent was unknown, but the remaining faults possessing the east-northeast trend were revealed for the first time by the Landsat imagery. Within the Mt. Alexander Block are several east to northeast trending lineaments which correspond to mapped faults. This trend is commonly developed throughout the study area in the form of major faults which appear to have undergone substantial movement. Probably there is a relationship between these major faults and the rotational axis of the folds throughout the study area (Plate 17).

The structure of the area is dominated by reverse faults that form the boundaries of numerous hilly ranges. Some major faults trend parallel to the fold axis and others cut diagonally across the axes (Plate 17).

On the enlarged Landsat images, three major rock units can be distinguished. The oldest unit crops out in the cores of the major anticlines and mountain ranges. These rocks are unstratified, indurated, characterized by a distinctive fine textured and rugged topography. Overlying the basement unit is a sequence of well-stratified rocks that crop out on the flanks of the anticlines. The stratification is expressed by variations in tone and resistance to weathering of the beds. Most of the strike and dip information on the map (Plate 17) is interpreted from outcrops of this well-stratified unit. The third rock unit consists of featureless alluvial and marine deposits on the footslopes.

Some further subdivision is possible. Plate 16 shows that the Cenozoic covering strata can be separated into two units on the basis of bedding, drainage system, resistance and tone. These features are not clear on a 1:1,000,000 scale MSS imagery which would not allow separation of these two units. Also three distinct strike ridges can be mapped off the Landsat images in the middle unit that separates rocks of different degradation rates.

7.9 CONCLUSIONS

Remote sensing provides valuable additional information for geological and geomorphological mapping and proved useful as a potential tool to delineate basic geomorphic units related to lithology and tectonic style. In this particular investigation standard and enhanced Landsat imageries were used after familiarity had been gained with the regional geology, following extensive field investigations, and as such, it provides an invaluable technique as part of an integrated research programme. If the most is to be gained from using the technique it is recommended that, whenever possible, Landsat studies should follow ground-based investigations as well as precede them.

The geomorphic features of a number of North Canterbury - South Marlborough rivers can be explained by an antecedent origin. The geomorphic features and the evolution of the fluvial system associated with active tectonic folding and faulting, such that diagnostic criteria for identifying such unstable areas, can provide constraints on an antecedent model. Major rivers took their present courses in an ongoing cycle of erosion introduced by gentle uplift and make their way in narrow gorges across uplifted mountain blocks or warping anticlines during the ongoing phases of deformation to which the present relief is due. Further evidence in support of the southward propagation model of the Marlborough Shear Zone comes from observation of the east-northeasterly trending lineaments recognised south of the Motunau Fault which seem to have originated most recently of all, and may well affect the future course of the Waimakariri River.

Most of the previously unmapped lineaments are the site of one or more epicenters and probably should be classified as active faults. The combination of geophysical data and remote sensing information is a promising approach to obtain quantitative evidence of the relative activity of neotectonic movements.

CHAPTER EIGHT

SUMMARY AND CONCLUSIONS

8.1 INTRODUCTION

One of the principal objectives of this study is to move towards developing a more comprehensive system of neotectonic mapping. Germane to the approach adopted, is the establishment of geomorphic models for the evolution of landscape features in different tectonic and climatic environments. In the case of this study the fundamental tectonic regime affecting North Canterbury appears to be a broad zone of deep seated dextral, or dextral-oblique, shearing expressed as a belt of en echelon folding in the Cretaceous and Tertiary cover rocks. The area is one of moderate rainfall and temperate climate, affected by the influence of late Pleistocene climatic variation producing periods of aggradation in the main river systems and fluctuating sea levels along the coast. Conventional mapping procedures used in basic geologic and structural studies have been employed, but the main emphasis has been to use geomorphological criteria to define tectonic activity. In particular, certain dynamic features such as the uplifting of marine terraces, the evolution of slope and fluvial drainage systems, tilting of fluvial terraces and the formation of late Quaternary deposits are seen to develop in response to the interaction of tectonism and climate.

It is evident from the outcome of this investigation that the evolution of slopes and drainage and the hydrological character of individual rivers are quite sensitive to the relatively slow processes of tectonic uplift and tilt. Apparently, even substantial rivers change their erosive and sediment-carrying capacities in response to changes in the rates and sites of tectonic activity, by crossing and recrossing hydrological thresholds. Such events appear to occur over much shorter periods of time than has been identified in any published accounts located by the writer.

Fundamental to the concepts used in compiling the Neotectonic map of the Lower Waipara (Plate 6) is the geomorphic model of drainage evolution on growing en echelon folds, summarised below and in Fig. 8.1.

8.2 A MODEL FOR THE GEOMORPHIC EVOLUTION OF EN ECHELON FOLDING

8.2.1 THE STRUCTURAL EVOLUTION OF EN ECHELON FOLDS

The geometry and genesis of en echelon folding has been investigated both empirically by the study of field examples and by means of model experiments. The latter are usually quite simple in form, consisting of a pair of rigid "basement" blocks which are set up to shear past each other and are covered with a variety of materials such as clay, silicone putty or even sheets of tissue paper, which will fold in response to the shortening strain generated parallel to the short axis of the ellipsoid of simple shear (e.g. Wilcox et.al, 1973; Harding, 1974; Graham, 1978; Cobbold and Quinquis 1980). Such experiments have limitations when translated to the realities of the natural environment. In particular the contact between rigid blocks and cover concentrates decollement along the contact and the "basement" cannot be involved in the fold process. Typically the folds tend to be dominated by canoe shaped anticlines and all vertical motion must be upwards because the design of the experiment inhibits the subsidence of synclines. In natural examples synclinal development is equally important and the detailed geometry of the folds is as affected by lithology as those produced in more simple compressional environments. Despite these limitations, certain features of the evolution of these folds do seem to translate to the natural case.

The first increment of shear strain will develop a component of shortening at 45^{o} to the shear direction and will begin to rotate clockwise or anticlockwise in the sense corresponding to the shear couple. The point at which visible folding will begin to develop in response to shortening will depend on the nature of the cover sequence. In most sedimentary covers, bedding anisotropy and the presence of relatively competent units will generate the necessary viscosity contrasts to promote the onset of flexure folding after shortening strain of a few percent has occurred. Thus the folds will begin to develop at a maximum angle of 45^{o} to the shear direction, perpendicular to the principal axis of shortening and will rotate as they grow in amplitude. If the folding is dominated by flexure, then there is a limit of about 30% shortening beyond which harmonic folding cannot continue to accomodate the shortening strain, at which point the axial trace is oriented at close to 30^{o} to the shear direction. Any further shortening forced on the central part of the fold must be taken up by limb thrusts and decollement. The experiments suggest that the strain is also disseminated by propagation of the folds along their lengths and by the generation of new folds such that the whole fold belt increases in width. The simple tissue paper simulation of this process by Graham suggests that there is a direct linear relationship between the width of the belt of folding and the finite angular shear strain. The disturbed belt thus extends substantially beyond the actual shear zone in the basement, a process exaggerated in these models by the ready detachment that occurs between the tissue paper and the surface beneath. Even so, there seems to be a length beyond which individual folds will not propagate. Complex local strain patterns must develop at these outer margins, where the particular mechanical constraints of the system limit the upward and outward dissemination of the shear strain. The experiments suggest that the characteristic sigmoidal shape of these folds develops, in part from the concentration of maximum rotation in the central part of the shear zone and also, from a relatively abrupt swing of the propagating tip as it approaches the outer limit. There the axial trace may curl back to angles as high as 70^{o} or 80^{o} to the shear direction.

The geometry and evident chronology of fold development in the Lower Waipara exhibits all these features and because the activity is young and ongoing, the modern drainage is profoundly affected by these processes. Any pre-existing river such as the Waipara, crossing such a shear zone, may be regarded as a material line which will undergo rotation, lengthening, or shortening, of the affected reach depending upon the initial angular intersection of the river course with the shear zone. It will be affected locally by the pattern of rising folds, all of which will modify bedform and gradient. Additionally, new local drainage systems will develop consequently on the evolving fold pattern with some distinctive characteristics, summarized below.

8.2.2 STAGES IN EVOLUTION OF DRAINAGE

Figure 8.1 is a generalized block diagram through an hypothetical emergent fold system. It attempts to illustrate the effect of active folding on the drainage pattern as interpreted from the lower Waipara area. Surface topography reflects the progressive deformation of a conformable sequence of marine strata in an en

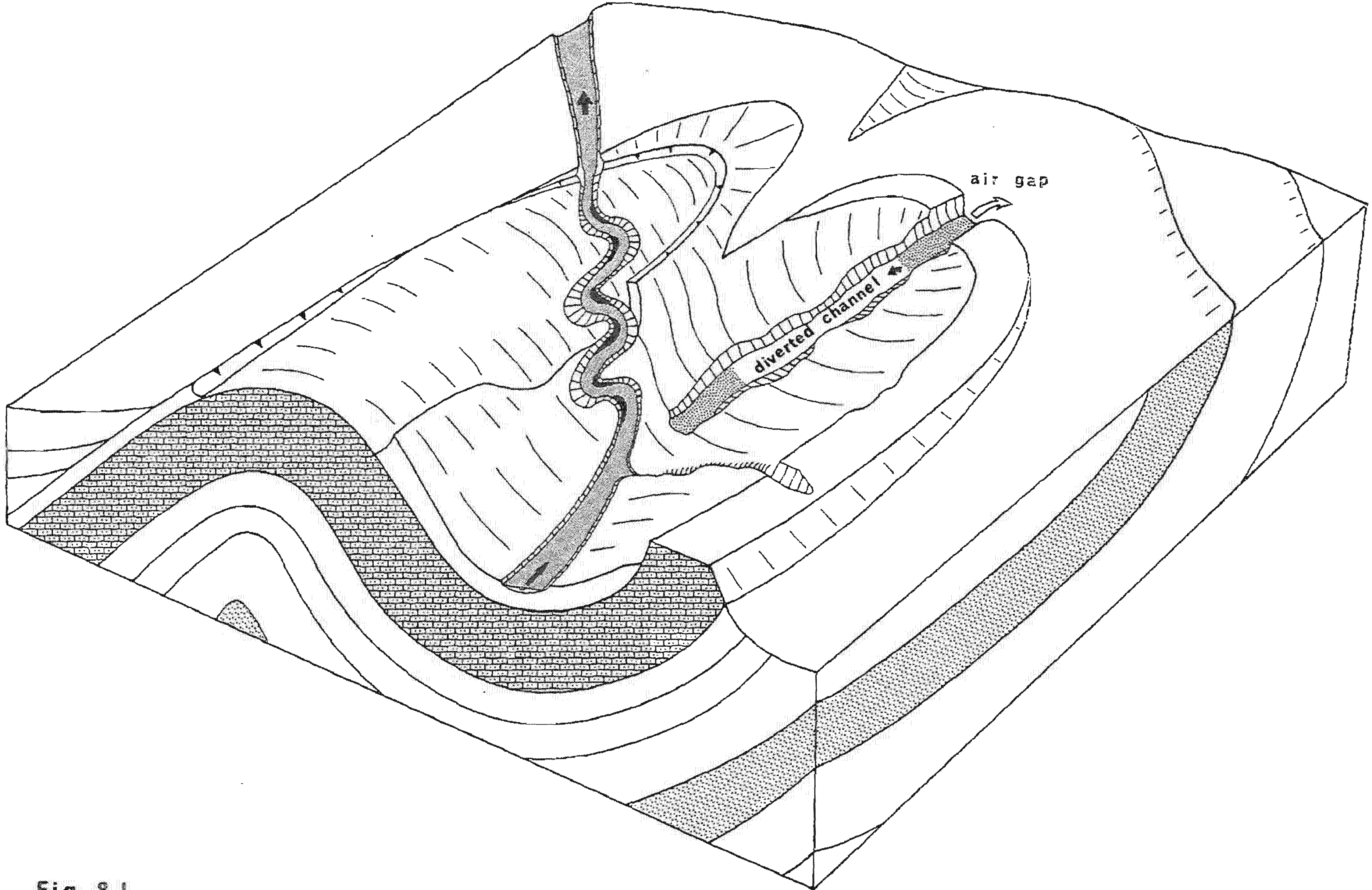


Fig. 8.1

Conceptual diagram of the effect of active folding on an alluvial drainage system echelon pattern of alternately plunging anticlines and synclines. In these young folds, the trough of a newly formed downfold, or syncline becomes a stream course. The crest of the adjacent anticline becomes a drainage divide where streams flow laterally down the anticlinal flanks. From this simple consequent pattern, in several of the catchments discussed, a consistent switch in the primary drainage channel occurs. The river is diverted across the adjacent anticline, through a broadly U-shaped upper valley and a steep inner gorge characterised by incised meanders on the upstream flank of the fold. An air gap is left in the syncline closure. It appears that the switching event may occur quite late in the development of the fold system. The chronology can often be linked to key regional geomorphic datum surfaces.

Certain aspects of these tectonically induced landscape features have implications which deserve further discussion.

1. Diverted channels and the mechanisms of antecedence

In all the examples considered, a reconstruction of the higher level channel positions indicates that there has been a significant realignment of the valley axes relative to their present positions. For major rivers, such as the Waipara, which cross the emergent belt of folding, the stream course is deflected as the plunging folds propagate across the system. Some of the divertionary processes may be gradual when slip-off occurs, as in the reaches above the gorge, but very often abrupt diversions from old to new channels occur, as in the relative alignments of the ancestral and modern valleys through the gorge.

The smaller streams occupying catchments within the fold belt are entirely controlled by the underlying structure from inception, and the diversion of the channel is clearly a consequence of backtilting and uplift in the synclinal hinge. The event occurs abruptly. At the time, the presence of the anticline crest presents less of a barrier than the syncline closure, even though the former may now lie at a greater elevation.

While the case of the Waipara River may represent an antecent river in the conventional sense, those that are diverted across the grain of the structure after the drainage pattern is already clearly established by the emergent folds, require a more forced, or active, process than the passive overprinting of pre-existing drainage on the structure. Semantically, "antecedent" is not a term which should still be applied to discordant drainage derived so late in the development of the structure.

Diversion by faulting is a well documented and readily understood process but folding is more difficult. Recently, the problem as been addressed by Oberlander (1985) when considering the very large scale cross-cutting of rivers entrenched across the fold belts forming major mountain ranges in the Zagros Mountains of Iran, the Himalayas, the Alps and the Appalachians. Two suggestions were made to account for the phenomenon.

1. In typical foreland fold belts, the folds grow sequentially away from the range front. Thus early formed folds will develop consequent drainage down the leading flank across the path of the next adjacent rising, but slightly younger, anticline. Thus the process is one of true antecedence.

2. In some cases, differential erosion of the anticline cores may lead to a process of stream capture that ultimately allows the whole drainage system to connect across the belt in a trellised system of rivers joined by headward erosion across the breached structural highs. This process may be enhanced by any cross folding or transecting structure which preferentially raises softer units to erosion level.

Neither of these hypotheses seem to be applicable here, and neither involve closure of the synclines as a diversionary mechanism. The relative elevation constraints at the time of diversion suggest that at some point in the shortening process the syncline closure occurs very rapidly, backtilting at a rate which cannot be matched by downcutting. At this point the nose of the adjacent anticline must be lower and behind the downstream end of the syncline. Further shortening is expressed as uplift and propagation of the anticline.

2. Channel position and transverse valley profiles

As uplift produced backtilting of the stream channels established across the rising anticlines, the initial response in the affected reach was of a marked transition to a meander form. With uplift accelerating to the point where stream power was insufficient to maintain lateral planation, the thalweg become fixed and incised, leaving relict strath surfaces preserved at higher levels. This sequence of events represents a shift in the system across two bedform thresholds and is governed by the balance of stream power to uplift and tilting. It is reflected in the broad and rounded upper levels of the valley sides and the inner, steep sided but terraced gorge (Fig. 8.2).

The association of incised meanders with one flank of an actively growing fold is well documented, but in these sinuosity studies (such as those of Adams,1980; Schumm et. al. 1982; Watson, 1982), the downvalley sides of the uplift axes have the higher values of sinuosity. These are all examples of intracontinental warping on a large scale and crossed by rivers of low initial gradients. Most seem to represent folding of low amplitude and the system has not evolved very far. In contrast, the North Canterbury rivers appear to represent systems that have evolved further on higher amplitude and shorter wave length folds superimposed on a gross regional uplift. This means that the basic hydrological characteristics of these rivers probably relate to much steeper background gradients (i.e. the gradient and corresponding bedform which the river would attain along its course, if this were controlled by regional uplift only).

The flume experiments which simulate the process (Schumm and Khan, 1972; Edgar, 1973; Ouchi, 1985 etc.) are set on specific gradients such that the basic (background bedform) is a braided channel. Downdip steepening of the gradient on the downstream flank of the simulated anticlines causes a transition across the threshold to a meander pattern. Conceptually, it seems likely that if the inital gradient were steep enough to form a straight, and even incised, channel this form would be accentuated on the downstream limb, but backtilting on the upstream limb would cause a reverse transition from straight to meandering. Coupled with this effect is the subsidence of the syncline relative to base level, causing aggradation and a transition to braided streams across the floor of the trough. Relative uplift of the anticline crest will cause incision, initally with the meanders capable of producing lateral planation of a broad valley, because of excess erosive power relative to uplift. The transition to deeply incised thalweg seems to be abrupt, and both the flume experiments and empirical observation of the natural systems suggests that channel morphology does not respond gradually to alteration of the parameters such as slope. The systems contain critical thresholds, below which the response will be negligible and above which response may be immediate and rapid. The precise point at which a threshold is crossed may be accelerated or delayed depending upon shorter term climatic influences which will impose periods of aggradation or degradation on the background system. This effect is likely to make quantification of the threshold transitions difficult.

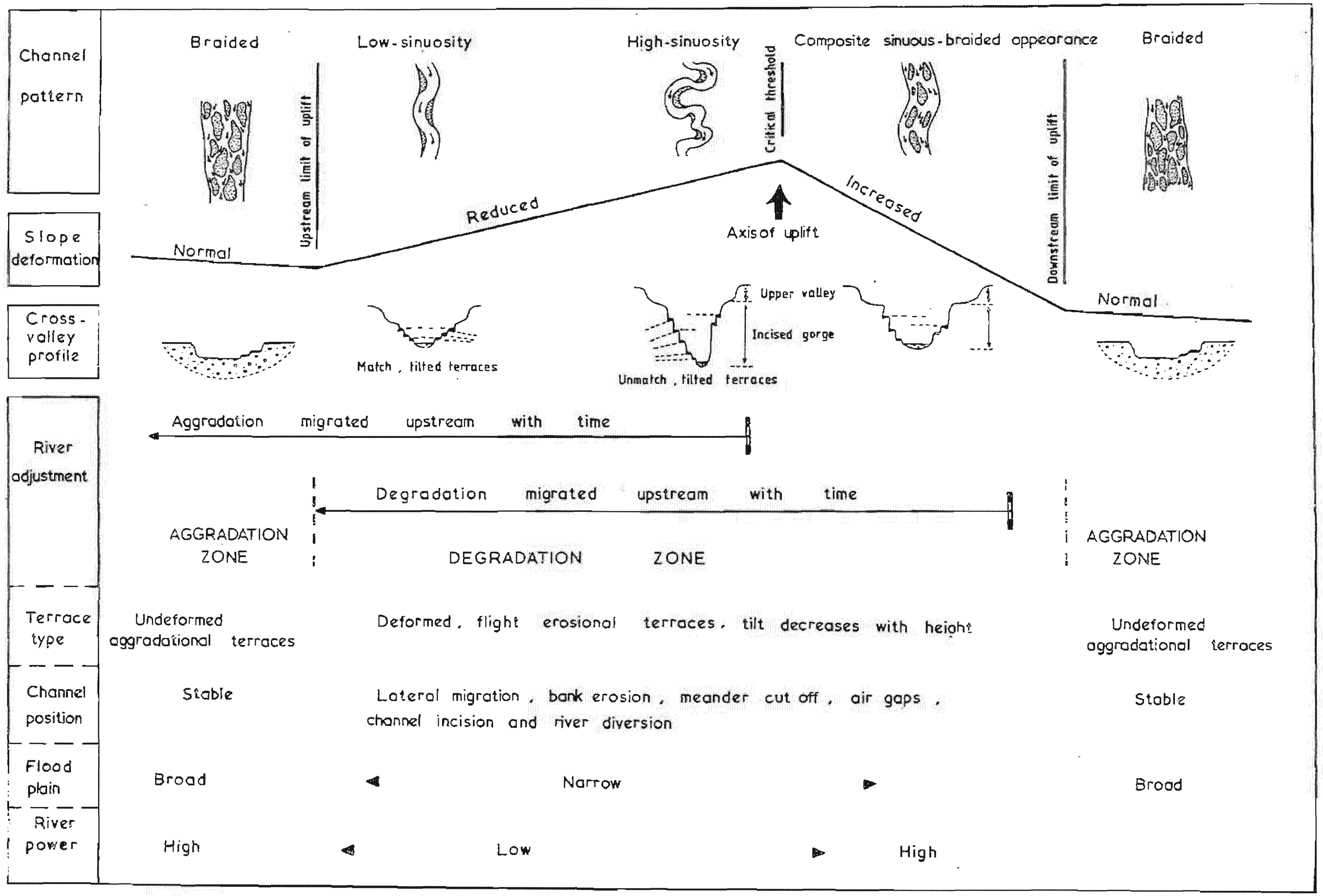
3. Terrace type

In the uplifted area, a sequence of flight, tilted matched and unmatched erosional terraces are formed, and the amount of tilt decreases with age (Fig. 8.2). Broad strath surfaces are most useful as synchronous datum surfaces because they represent sustained periods of equilibrium. In this study there are excellent examples of ancestral and modern straths that probably continue downstream for many kilometres.

4. Long term river adjustment

There is a transition from aggradation within the synclines to degradation in the reaches breaching the anticline. During the main phase of accelerated uplift of the emergent anticline, the threshold transition point between the two modes will migrate outward and downward so that at higher levels degradational terraces occupy only

Fig. 8.2 Adjustment of a braided river to anticlinal uplift across it



the central part of the gorge near the anticline crest. Remnants of strongly tilted aggradation terraces may be found high on the flanks, as in the case of the Teviotdale Surface on the Cass Anticline. Lower down, aggradation terraces, even when climatically induced may not be found far downstream of the entrance to the gorge as in the case of the Canterbury Surface. If uplift slows, ultimately the river will cause long term adjustments in which lateral planation and a braided bedform will work back into the lower end of the gorge, while downcutting and the production of degradation terraces will migrate back from the upper end of the gorge cutting into the upstream trough fill (Fig. 8.2).

The pattern of the lower Waipara River changes in appearance from braided, through low to high sinuosity, to composite sinuous braided and finally back to braided. These changes took place at critical threshold values of stream power, channel gradient and sediment load in response to valley slope deformation. While this succession is fundamentally a reflection of accelerating uplift across the Cass Anticline, the most recent threshold change is a cessation of simple incision across the fold and the beginning of lateral planation of a new strath surface covered by a bedload veneer. It seems likely that any future shortening will be concentrated on the limb thrust formed by the Hamilton Fault.

5. Effects of faulting

While this study is concerned primarily with folding, faulting is significant in modifying the fold induced processes. Thrust faulting of the limbs of locked flexure folds after the limit of shortening is attained will permit uplift to continue, but the anomalies in gradient and related effects will clearly become concentrated at the fault outcrop. Local backtilting can occur, even on the downstream anticline flank. In a fundamentally shear regime, strike-slip faulting may be anticipated, particularly in the form of short reidel shears either singly or in en echelon swarms oblique to the folds. Possibly an example of one of these may have been instrumental in the diversion of the Waipara River from its old course down the lower reaches of the Teviotdale River to its present outlet.

The Rakaia Gorge case study is not strictly a contribution to the basis for the fold model, because the deformation there is dominated by upstream backtilting induced by a major reverse fault. It was included because it is the most readily available example of a much larger antecedent river, providing a means of considering the sensitivity to tectonic processes of a river of greater power. The backtilting does compare with the effects of the upstream dip reversal of an anticline, again producing deeply incised meanders which increase in amplitude up to the fault trace. Here the transition back to braided bedform occurs abruptly. Uplift and tilt are clearly rapid enough to induce threshold transitions in the main river and divert small tributaries as in the case of Boundary Stream.

8.3. UPLIFT RATES

It is clear from the qualitative evidence of the threshold tansitions just discussed and recogised in the Lower Waipara landscape, that uplift rates vary from one place to another because of the differential effects of folding. It is also evident that rates at specific sites have varied within the timespan of the Otiran Glaciation and the Holocene i.e. the rates can be variable within periods of the order of 100Ky. At some localities it has been possible to quantify estimates of uplift rates where datable material, or surfaces to which ages can be assigned by correlation, can be tied stratigraphically to a measurable vertical displacement. Average values of around 2 mm/yr represent the order of magnitude of the uplifts being experienced in this area but it is evident that in some cases, the data at individual sites represents total uplift (as in the case of marine terrace uplift) and at others, differential rates due to folding only. It is clear from the geomorphic analysis and from considerations of the derivation of the measurements, that the deformation at any locality will be affected by processes which can be separated into four orders.

8.3.1. FIRST ORDER REGIONAL UPLIFT

The shear zone around the northern margin of the Canterbury Plains clearly separates areas of much higher average relief in the hinterland from the coastal plains and it seems likely that regional emergence of the whole fold belt is taking place. This uplift will be responsible for setting the background gradient on the major rivers having their source well back into the foothills of the Southern Alps, as in the example of the Waipara River, even though this gradient may be highly modified by the two structural barriers through which it is incised.

8.3.2. SECOND ORDER DIFFERENTIAL UPLIFT DUE TO FOLDING

One of the evident consequences of the geometry of folding in general is the non-linear relationship between shortening strain and uplift. The amplitude of a flexure fold grows slowly at first then accelerates rapidly to tail off as the simple flexure shortening limit is reached. The form of this curve becomes flatter and with longer tails, as the viscosity contrast within the units decreases and the fold form departs from a simple parallel style. When the shortening is generated as a consequence of shear strain, the relationship between uplift rate and the primary driving system is further complicated by the non-linear relationship between the rate of shear strain and the shortening rate. While this effect will modify the uplift rate expressed as the growth in amplitude, it will not affect the basic cycle of acceleration and deceleration. Thus any antecedent river crossing such a fold should show a basic pattern of broad upper valley cut during early uplift, an incised inner gorge once the critical uplift/stream power ratio is reached and a final period of renewed lateral planation as the fold approaches the limit of closure. Such a pattern should be regarded as normal and does not necessarily imply changes in the fundamental level of tectonic activity, although the variables in the system would make a genuine change in the rate of horizontal shear very difficult to detect.

8.3.3 THIRD ORDER FLUCTUATIONS IN THE RATE OF UPLIFT AND THE POSSIBLE CAUSES

From the foregoing discussion it is evident that during the normal progress of folding induced by shortening, the rate of differential uplift will pass through a cycle of acceleration and deceleration with time. The precise form of the cycle will depend upon the mechanical constraints induced by lithology on the fold process and in the case of shortening, as a secondary response to shear, the shortening rate will be non-linear. However, it is clear from the general evidence of steep dips in bedding beneath the strath surfaces, even when these are markedly tilted, that in the Waipara, Carrington, Yellow Rose, Kate and Teviotdale Gorges, considerable shortening strain must have already taken place prior to that portion of the fold history documented by geomorpholocical evidence of fluctuations in uplift rate. In each case the entrapment of the river across the rising anticline is followed by relatively slow uplift, accelerating to entrench the stream. In detail, the pattern of events differs as does the timing of stage transitions so that each system is responding to local effects in the growth of the Cass, Black and Kate Anticlines respectively.

This cycle occurs in the Waipara from a period preceding the 80,000 year old marine terrace when the ancestral course laid fluvial gravels across this surface, followed by the aggradation of the Teviotdale Surface, diversion of the river and rapid incision of the inner gorge. Apparently warping ceased by about 5,000 years ago (Chapter 6.3). In Carrington Creek, diversion of the creek across the Black Anticline post-dates the Teviotdale Surface, and the downcutting process is interrupted by the overriding aggradation of the Canterbury Surface onto a broad strath terrace, indicating that the system had not yet entered the rapid incision stage (Chapter 6.4). Both the Cass and Black Anticlines have deeply eroded cores when traced up plunge from these two gorges, indicating that the central parts of these folds have been uplifted for some time. The Kate Anticline is clearly topgraphically younger, being expressed directly as a rounded surface with little modification, although it must have been already in existence before being breached at both ends by the Teviotdale River and Kate Creek respectively.

Kate Creek occupied the axis of the Teviotdale Syncline during the period of the 125,000 to 80,000 year old high sea level stands and was diverted across the Kate Anticline after the youngest of the related marine terraces. Since then downcutting has continued with active upstream migration of a distinctive knickpoint. The Teviotdale River has a similar chronology (Chapter 6.5).

These observations suggest that there are localised processes that produce short term changes in uplift rate. From the evidence outlined, the origin is not clear. It is likely that the highly complex strains developed around the propagating tip of a fold may account for the relative rapidity with which events succeed each other. Certainly, in simple model experiments, the period of time over which the steeply plunging fold tip begins to grow and tightens to the limit of flexure at any one point is very quick, relative to the growth of the whole fold structure and to the increment of shear strain in the underlying shear zone over that period. Inhomogeneity of strain in the basement with transfer of deformation from one fault to another is probable and may be a contributing factor. There does not seem to be a strong case to suggest that these short term effects are sufficiently synchronous throughout the area to be attributed to fundamental changes in the rate of regional shear strain.

8.3.4 FOURTH ORDER PETURBATIONS AND THE QUESTION OF COSEISMIC DEFORMATION

Unequivocal evidence of past seismic activity apart from fault displacement of geomorphic surfaces, was not found. The radiocarbon dated (1150 +/- 55 Yr.B.P.) modification to terraces induced by faulting in Yellow Rose Creek being the only directly dated event, others being inferred from the stratigraphic position of the surfaces being displaced. On these criteria faults have been distinguished on the Neotectonic Map according to a classification scheme which is slightly modified from the one currently adopted by the New Zealand Geological Survey (Plate 6).

1. Highly active:

CLASS IA - Faults which have repeated movement between 1,000 and 10,000 years ago.

CLASS IB - Faults which have moved once between 1,000 and 10,000 years ago. (In the study area no displacements younger than 1,000 years ago were identified, but clearly faults on which there is evidence of more recent activity could either be categorised separately, or these classes extended to include them.)

2. Moderately active:

CLASS II - Faults which have moved once between 10,000 and 50,000 years ago but this category may also include some Class IB faults for which there is evidence of a history of movement prior to a displacement event within the last 10,000 years.

3. Inactive:

CLASS III - Faults which have moved last between 50,000 and 500,000 years ago.

The coincidence of a small strike slip fault with the saddle marking the location at which the lower course of the Waipara was diverted from the lower Teviotdale valley to its present position, suggests that activity on this fault may have been instrumental in the diversion process. Possibly the moment of diversion may have been associated with coseismic movement on the fault, but the topographic setting which allowed the spill over to take place was a consequence of cumulative tilting of the old Teviotdale Surface. This site also coincides with the trend of the Kate Anticline and migration of the fold nose may account for several topographic anomalies recognised on the surface of the ancestral Waipara Valley in this vicinity.

Landslides are common throughout the area but many are superficial and probably climatically induced. Large and deep seated failures such as the Montserrat slide have an ongoing history of movement so that any evidence of the initating event has been destroyed, consequently no attempt was made to estimate paleoseismicity from landsliding.

The question of whether or not part, or all of the folding is coseismic remains speculative. If the interpretation of the mechanism of folding is correct, then it seems likely that they are underlain by an anastomosing fault system in the basement, the continuation of which can be seen exposed at the surface along the strike of the zone in the Porters Pass area, where it is currently under study. There, a growing body of evidence supports the view that the zone is seismically active although evidence of the recurrence intervals has not yet been fully established.

8.4 TOWARDS DEVELOPING A SYSTEM OF COMPREHENSIVE NEOTECTONIC MAPPING

8.4.1 OBJECTIVES

Traditionallly, in New Zealand and largely elsewhere, the representation of neotectonic deformation has been confined to identification of active faults and the documentation of dis placement data and dates of offsets. Classification schemes of relative activity have been proposed based on these obs ervations and on some broader geomorphic criteria, as for example Bull and McFadden (1977) categorisation of range bounding faults in arid environments. Attempts have also been made to contour uplift rates and represent components of tilting, shear strain and shortening, derived mainly from geodetic data.

It seems reasonable to suggest that maps could be more informative if the means of representing much more information on the nature of ground deformation could be conveyed. There is a need to represent the nature and geometry of deformation being generated by a wide range of structures in a variety of tectonic environments. The observed components of deformation derived from either geophysical or geomorphic criteria should be related to geometry of the underlying structures, the mechanics of which place some constraints on the interpretation of these data and the way in which observed rates may be extrapolated.

There is also a need to represent aspects of the past history so that short term e.g. geodetic observations can be compared to longer term displacements derived from geomorphic observation and to document sites at which components of the strain are accelerating or decelerating. Ideally it would be useful to show any migration of activity or change in its character within an appropriate time interval. Clearly there are many problems both cartographic and in state of the art before such objectives could be fully realised.

These general objectives lie behind the design of the neotectonic map derived from this study (Plate 6). The general methodology adopted is summarised below.

The character of active deformation in North Canterbury was approached at two levels:

1. A regional analysis derived largely from remote sensing and existing geological maps, and

2. Local mapping of areas of particular interest using aerial photographs and ground based observation and surveying.

8.4.2. REGIONAL ANALYSIS

The majority of the maps representing the results of the general analysis of the North Canterbury area are derived from the available satellite multispectral imagery which unfortunately did not include either the very recently available SPOT imagery or any Side Scanning Radar. Both these latter have considerable potential. Other geophysical data are also useful to correlate with observed lineaments and major geological features, most useful in this case being earthquake epicentre locations.

Recent studies of Landsat imagery have shown that major and minor lineaments, probably of deep-seated origin, traverse areas not normally thought of as being tectonically active. Remote sensing, combined with traditional methods of geommorphic analysis, supply the techniques by which a qualitative interpretation can be made. Clearly, detailed quantitative analysis of the landscape relationships as a means of assessing the level of activity will be difficult and remains a future objective in improving significantly our understanding of the long-term evolution of landscapes in such tectonically active regions.

Several companion maps were prepared to a common scale (Plate 15 & Figs. 7.5 to 7.9). These serve to delineate the main geological provinces and identify major lineaments and to tie these to the texture and pattern of drainage . Noted in particular are the sites of drainage pattern anomalies that may indicate active deformation. Significant correlation of these anaomalies with rather subtle lineaments across the Canterbury Plains was observed, where atypical zones of meandering mark the point where the streams cross the lineaments and which also correlate well with indications of seismic activity. There is good evidence of active folding in a belt around the margin of the Plains indicating that a continuation of the Lower Waipara style of deformation reflects a broad zone of shearing extending southwestward to the Porters Pass area. Many rivers are entrenched into antecedent gorges north and west of this belt across a pattern of fault bounded and probably warped blocks.

Wherever possible, mapping procedures from remotely sensed data should utilize all available imagery, including active and passive multiband sensors of varying scales. For example, seven single-band images, as well as enhanced colour composites, were compared. Band 7 was found to be the single most useful spectral band for geomorphic mapping. Rivers and drainage patterns show clearly and help to emphasise the pattern of hills and valleys. Several distincitive drainage patterns can be seen (Fig. 7.5). Enhanced image processsing of Landsat images reveals significantly more information than the standard product photographic versions of the same images. This is ill ustrated by comparisons between both types of images as used in the photogeological interpretation of the morphotectonics of North Canterbury.

8.4.3 LARGER SCALE MAPPING OF LOCAL AREAS

The methodology here is pertinent to the setting ofthe Waipara area and relates to fluvial and coastal marine processes, but the general approach should be of wider application. The basic procedure is summarised in the following steps.

A. General principles

1. The landscape elements are delineated by conventional geomorphic mapping techniques and grouped according to those associated with a chronology of events separated by threshold changes in each catchment.

2. The influence of differential erosion of bedrock controlled structures and the changes induced by climatic variations have to be filtered out from the effects of tectonism. The indexing of stream gradients using computer analysis of data digitised from existing topographic maps was found to be a particularly useful technique. The mapped positions of anomalies in the gradients of not only the major drainage channels, but also the minor tributaries, were compared with both the basic geological and geomorphic maps. Those anomalies associated with bedrock control and base level changes induced by sea level change along the coast could be readily identified. This left a residue of breaks in gradient which were tied to both known and previously unidentified faults and to the effects of warping and uplift.

3. The information is not available to attain total quantification of the evident deformation but the pattern of threshold transitions provides a feel for the general level of relative activity. Crucial to a partial quantification, is the identification

of key datum surfaces for which general ages can be assigned on the basis of the regional Pleistocene chronology, in this case the two main aggradation surfaces the Teviotdale (circa 70 Ky) and Canterbury (circa 10-14 Ky) Surfaces and the succession of marine terraces.

4. "Spot" quantification of one or more deformation parameters (uplift or vertical offset in this area) can be added to this general pattern. This is possible wherever the fortuitous association of datable material, or a datum line or surface of known age, can be related stratigraphically to a site where the displacement component can be measured.

B. Marine terraces as datum surfaces

Marine terraces have the fundamental advantage of original horizontality and can be related to the global datum of sea level. Since the dating of the New Guinea sequence of uplifted reefs on the Huon Peninsula (Chappell, 1974) and subsequent studies which have established a chronology, studies of the deformation of marine strandlines have become a standard technique in neotectonic studies. Few such shorelines can be dated directly and most quantification depends upon the reliability with which the age of a surface can be established. The coastal terrace system studied here illustrates most of the typical problems encountered in this type of study and because there is good control on the geometry of the fold structures and some independent data on the deformation chronology, this example suggests that some common practices may be based on assumptions that are too simplistic.

The main problem is the unequivocal dating of surfaces. Most appropriate direct dating methods such as amino acid racemisation and uranium series radiometric dating of shell material are laborious, expensive and the facilities not always available. Tephra deposits on the surface give a minimum age where these can be found, but depend on either direct fission track dating or chemical fingerprinting for identification. Consequently many chronologies rely on relative altitude to estimate ages. The basic premise is usually an assumption of uniformity of uplift rate. Once the correct age and sea level correction factor has been allocated to one terrace elevations of all the others which should be present for a given uplift rate may be predicted from the New Guinea chronology and compared to the observed elevations, such that they plot on a uniform gradient when elevation is plotted against assumed age (e.g. Bull and Cooper, 1986).

This study suggests that in an environment of active folding, uniformity of uplift rate cannot be assumed. Despite this constraint, elevations are still almost the only data available. However, there are key stratigraphic relationships where the marine terrace sequence can be tied to episodes of aggradation correlated with major stadials of the Otiran Glaciation that impose requirements of internal consistency in allocating terraces to particular interglacial or interstadial high stands of sea level, as detailed in Chapter 5.7.

Precision in determining the elevation of a marine planation surface is limited by the thickness of the littoral and regolith deposits on it. Ideally the nickpoint at the back of the cut surface should be used, limiting data points to exposures, usually in cross-cutting gullies.

Although the surfaces may be quite broad, the fact that only this knickpoint relates precisely to sea level means that each terraces provides only a line imprinted across the structure, limiting the amount of three dimensional information on deformation geometry which can be determined. Each strandline occupies a different part of the structure so that elevation differences are not strictly comparable measures of uplift at a single site. The isobase technique of Ghani (1978), Pillans (1983), Bishop (1985) is of doubtful validity in a tectonic environment of complex geometry, because there is no reason to assume a fixed hinge line and the sinuosity of the axial trace presents problems and too many degrees of freedom in projecting the data.

8.5 CONTRIBUTIONS OF THIS STUDY TO CURRENT VIEWS ON THE MARLBOROUGH SHEAR ZONE

The interpretation of the Lower Waipara area as one of active shear folding, and the general projection of this zone in the regional study southwestward as a substantial tectonic boundary, is compatible with Carter and Carter's (1982) suggestion that this belt marks the southern limit of the Marlborogh Shear Zone and the current plate boundary. However, the zone is clearly more complex and extends southward further than simply equating it with the Motunau and Porters Pass Faults.

Walcott (1984) in his analysis of existing geodetic information carried his data as far south as the Lower Waipara, where shear strain was shown to be decreasing from a maximum in excess of 0.4 microradians/yr in the central part of the Shear Zone to a zero contour passing through the Canterbury Plains and Lower Waipara. This does not seem to be very compatible with the evidence of this study and is better supported by the more recent analysis of Reilly (1987).

Reilly's data is carried further south to include the Canterbury Plains and computes shear strain values in the range of 0.2 microradians/yr over the Plains area. Contours of dextral rotation trend northwest to southeast with a maximum value of -0.6 microradians/yr near Christchurch to -0.4 at Waipara. These data support the interpetation of this study, that there is active deep-seated shearing along lineaments beneath the Plains, a possibility that has obvious implications for seismic hazard potential in the metropolitan area.

8.6 SUGGESTIONS FOR FUTURE RESEARCH RELATING TO THIS STUDY

The results of this study and the methods and principles developed, suggest some potential for extensions of this work along a number of lines.

1. There are a number of sites around Canterbury where folding of this style suggests that the geomorphic model developed here could be used to produce similar maps locally, for example the Ashley and Selwyn areas. Shear folding is widely documented from other tectonically active areas and it would be of interest to test the universality of the model, particularly under other climatic conditions and rivers of greater erosive capacity.

2. Systematic studies could be made of areas undergoing deformation of quite different tectonic style, initially choosing locations that are as typical as possible, in order to develop diagnostic geomorphic criteria based on the threshold principle.

3. Physical modelling involving an extension of the flume based experiments of Ouchi and others might be attempted, to test firstly the effect of fundamental gradient on the bedforms developed over

folds, and secondly to try to emulate shear folding.

4. Computer based modelling of actively growing folds in profile could be used to explore and quantify stream gradient index patterns for streams of variable erosive capacity. There is a potential for extracting more information from stream gradient data than has been possible in this study.

5. The principles and mapping techniques adopted in this study have direct application to planning procedures and can be incorporated into zoning maps and hazard assesment investigations.

8.7 CONCLUDING STATEMENT

Despite the extensive literature which has appeared in recent years on modern ground deformation and neotectonics, research in this area is still under-subscribed and is still at the stage where basic principles and methodology are being established. The full potential as a research discipline has yet to be realised. With recent technological advances and ever increasing coverage, air photo and remote sensing techniques are ideally suited to addressing this topic, particularly in tectonically highly active areas. Through remote sensing studies it may be possible to advance further our ideas on the neotectonic behaviour of the Earth's crust.

ACKNOWLEDGEMENTS

I wish to express my sincere thanks to Mrs J.K. Campbell who has done more than her duty as supervisor, and also to Professor A.R. Crawford and Dr J. R. Pettinga who co-supervised the study and for constructive criticism of the manuscript. I am also grateful to the Iraqi Government and the University of Mosul for the financial support of my scholarship. It is a pleasure to acknowledge the great help from all the staff of the Geology Department and the Geological Survey; the following in particular deserve special mention: Dr D.I. Mackinnon, Mr S. Nathan, Mr G.H. Browne, Mrs W. Nuthall and D.J. Jones.

Mr A.H. Downing contributed valued technical assistance with photography.

Assistance in the field from Alister McCrae and Gordon McLean was greatly appreciated and the use of survey equipment belonging to the Department of Civil Engineering is gratefully acknowledged. Assistance in the computer programme from Dr B. Hockey and Mr D. Mark was greatly appreciated and the use of the digitizer equipment belonging to the School of Forestry, University of Canterbury, is gratefully acknowledged.

Thanks go to Mrs S. Tye for the speed and efficiency with which she typed the text. Thanks go to various landowners in the area, for their generous hospitality and permission to work on their land, in particular to Mr and Mrs Nesbit of Tiromoana, and Mr and Mrs Greenwood of Teviotdale.

Thanks also to many colleagues for their collaboration and stimulating discussions on various aspects of tectonic geomorphology, especially to Professor J. Soons, Department of Geography, University of

Canterbury, Professor W. B. Bull, Department of Geoscience, University of Arizona and Professor S.A. Schumm, Department of Earth Resources, Colorado State University during their visits to the study area.

Special thanks to my parents, family, friends, and my daughter Rawa, without whose encouragement and faith, the project would not have been possible.

Finally, may I express a word of gratitude to my wife, Sana Kamal Abdul Hamed, who has had to live with this thesis for the best part of five years.

REFERENCES

ADAMS, C.J.D.; OLIVER, P.J. (1979): Potassium argon dating of Mt Somers Volcanics, South Island, New Zealand. N.Z. Journal of Geology and Geophysics, 22, 455-463p.

ADAMS, J. (1980a): Contemporary uplift and erosion of the Southern Alps, New Zealand. Geological Society of America Bulletin, part II, v.91, 114p.

ADAMS, J. (1980b): Active tilting of the United States midcontinent geodetic and geomorphic evidence. Geology, 8, 442-446p.

ADKIN, G.L. (1930): Origin of the Manawatu Gorge. N.Z. Journal of Science and Technology, 11, 353-356p.

ANDREWS, P.B. (1963): Stratigraphic nomenclature of the Omihi and Waikari Formations, North Cnaterbury. N.Z. Journal of Geology and Geophysics, 6, 228-256p.

ANDREWS, P.B. (1968): Patterns of sedimentation during early Otaian (Early Miocene) time in North Canterbury, New Zealand. N.Z. Journal of Geology and Geophysics, 11, 711-752p.

ANDREWS, P.B. and others. (1976): Lithological and paleontological content of the Carboniferous-Jurassic Canterbury Suite, South Island, New Zealand. N.Z. Journal of Geology and Geophysics, 19, 791-810p.

ANDREWS, P.B. and others. (in prep.): Lithostratigraphic nomenclature for the Upper Cretaceous and Tertiary sequence of central Canterbury, South Island. N.Z. Geological Survey Record.

ARABASZ. W.J.; ROBINSON, R. (1976): Microseismicity and geologic structure in the northern South Island, New Zealand. N.Z. Journal of Geology and Geophysics, 19, 569-601p.

BERRYMAN, K.R. and others. (1987): Evolution of Holocene coastal geomorphology at Pakarae River, eastern North Island, New Zealand. In Ota, Y. (ed), Holocene Coastal Tectonics of Eastern North Island, 23-33p.

BERRYMAN, K.R. (1980): Late Quaternary movement on White Creek Fault, South Island, New Zealand. N.Z. Journal of Geology and Geophysics, 23, 93-101p.

BERRYMAN, K.R. (1979): Active faulting and derived PHS directions in the South Island, New Zealand. Royal Society of New Zealand Bulletin, 18, 29-34p.

BERTHE, D.; BURN, J.P. (1980): Evolution of folds during progressive shear in the South Armorican Shear Zone, France. Journal of Structural Geology, 2, 127-133p.

BIBBY, H.M. (1981): Geodetically determined strain rates across the southern end of the Taupo-Kermadec-Hikurangi subduction zone. Geophysical Journal R. astr. Society, 66, 513-533p.

BISHOP, D.G. (1985): Inferred uplift rates from raised marine surfaces, Southern Fiordland, New Zealand. New Zealand Journal of Geology and Geophysics, 28, 243-251p.

BLOOM, A.L. and others. (1974): Quaternary sea level fluctuations on a tectonic coast: new 230th/234u dates from the Huon Peninsula, New Guinea. Quaternary Research, 4, 185-205p.

BRADSHAW, J.D. (1975): The folds at Castle Hill (Canterbury) and their bearing on Kaikouran deformation style in the Canterbury Basin. Journal of the Royal Society of New Zealand, 5(2), 209-217p.

BRADSHAW, J.D.; NEWMAN, J. (1979): Low-angle thrusts in Cenozoic rocks in Canterbury, New Zealand. N.Z. Journal of Geology and Geophysics, 22, 435-442p.

BRADSHAW, J.D. and others. (1980): Carboniferous to Cretaceous on the Pacific margin of Gondwana. Fifth International Gondwana Symposium, Wellington, New Zealand. M.M. Cresswell & P. Vella (eds), 217-221p.

BROME, G.H.; FIELD, B.D. (1985): The lithostratigraphy of late Cretaceous to early Pleistocene rocks of North Canterbury, New Zealand. N.Z. Geological Survey, Record, 6, 63pp.

BUCKNAM, R.C.; ANDERSON, R.E. (1979): Estimation of fault-scarp ages from a scarp-height-slope-angle relationship. Geology, 7, 11-14p.

BULL, W.B. (1984): Tectonic geomorphology. Journal of Geological Education, 32, 310-324p.

BULL, W.B. (1979): Threshold of critical power in streams. Geological Society of America, Bulletin, 90, 453-464p.

BULL, W.B.; McFADDEN, L.D. (1977): Tectonic geomorphology north and south of the Garlock fault, California. In Doehring, D.O. (ed), Geomorphology in Arid Regions. Binghampton, Suny, 115-138p.

BULL, W.B.; KNUEPFER, P.L.K. (1987): Adjustments by the Charwell River, New Zealand, to uplift and climatic changes. Geomorphology, 1, 15-32p.

BULL, W.B.; COOPER, A.F. (1986): Uplifted marine terraces along the Alpine Fault, New Zealand. Science, 234, 1225-1228p.

BURENT, A.W.; SCHUMM, S.A. (1983): Alluvial-river response to neotectonic deformation in Louisiana and Mississippi. Science, 222, 49-50p.

BURROWS, C.J. (1979): A chronology for cool-climate episodes in the Southern Hemisphere 12,000-1000 yr B.P. Paleogeography, Paleoclimatology, Paleoecology, 27, 287-347p.

BURROWS, C.J. (1983): Radiocarbon dates from Late Quaternary deposits in the Cass District, Canterbury, New Zealand. N.Z. Journal of Botany, 21, 443-454p.

BUSK, H.G. (1929): Earth flexures. C.U.P., 106p.

BUTAKOV, G.P. (1970): Experiment in the mathematical description of the form of longitudinal river profiles. Geomorphology (transl. from Geomorfologiya), 1, 336-340p.

CAMPBELL, J.K.; YOUSIF, H.S. (1985): Tectonic geomorphology of the lower Waipara gorge, North Canterbury. Geological Society of New Zealand, Christchurch Conference, Field Trip Guides 32b, 53-69p.

CARR, M.J.: The stratigraphy and chronology of the Hawera Series. Marginal succession of the North Canterbury Coast. Christchurch, University of Canterbury, 1970. (Thesis: Ph.D.: Geology).

CARRYER, S.J. (1967): The glacial deposits along the northern flank of the Mount Hutt Range. N.Z. Journal of Geology and Geophysics, 10, 1136-1144p.

CARSON, M.A. (1984a): Observations on the meandering-braided river transition, the Canterbury Plains, New Zealand: Part 1, 12-17p.

CARSON, M.A. (1984b): Observations on the meandering-braided river transition, the Canterbury Plains, New Zealand: Part 2, 89-99p.

CARTER, R.M. and others. (1982): Oligocene unconformities and nodular phosphate-hardground horizons in western Southland and northern West Coast. Journal of the Royal Society of New Zealand, 12, 11-46p.

CARTER, R.M.; CARTER. L. (1982): The Motunau Fault and other structures at the southern edge of the Australian-Pacific plate boundary, offshore Marlborough, New Zealand. Tectonophysics, 88, 133-159p.

CHAPPELL, J. (1983): A revised sea-level record for the last 300 000 years from Papua New Guinea. Search, 14, 99-101p.

CHAPPELL, J.(1975): Upper Quaternary warping and uplift rates in the Bay of Plenty and West Coast, North Island, New Zealand. N.Z. Journal of Geology and Geophysics., 18, 129-155p.

CHAPPELL, J. (1974): Geology of coastal terraces, Huon Peninsula, New Guinea, a study of Quaternary tectonic movements and sea level changes. Geological Society of America, Bulletin, 85, 553-570p.

CLAYTON, L. (1966): Tectonic depressions along the Hope Fault a transcurrent fault in North Canterbury, New Zealand. N.Z. Journal of Geology and Geophysics, 22, 535-553p.

COATES, D. (1972): Environmental geomorphology and landscape conservation. Volumes I, II, III. Wiley.

COBBOLD, P.R. AND QUINQUIS, H. (1980): Development of sheath folds in shear regimes. Journal Structural Geology, 2, 119-126P.

COLE, J.W.; LEWIS, K.B. (1981): Evolution of the Taupo-Hikurangi subduction system. Tectonophysics, 72, 1-21p.

COLLINS, B.W.: The geology of the eastern Omihi District. Christchurch, University of Canterbury, 1939. (Thesis: M.Sc.: Geology).

COLMAN, S.M. (1983): Progressive changes in the morphology of fluvial terraces and scarps along the Rappahannock River, Virginia. Earth Surface Processes and Landforms, 8, 201-212p.

COOKE, R.V. (1974): The rainfall context of Arroyos initiation in southern Arizona. Zeitschrift fur Geomorphologie, Suppl. 21, 63-75p.

COOKE, R.V.; REEVES, R.W. (1976): Arroyos and environmental change in the American Southwest. OUP.

COTTON, C.A. (1913): Physiography of the Middle Clarence Valley. Journal of Geography, 42, 225-246p.

COTTON, C.A. (1958): Geomorphology. An introduction to the study of landforms. 7th ed. Whitcombe and Tombs Ltd, 505p.

COTTON, C.A. (1941): Landscape, as developed by the processes of normal erosion. C.U.P., 301p.

COX, P.T. (1926): Geology of the Rakaia Gorge District. Transactions of the New Zealand Institute, 56, 91-111p.

CRITTENDEN, R.M.; MUHS, D.R. (1986): Clif-height and slope-angle relationships in a chronosequence of Quaternary marine terrace, San Clemente Island California. Z. Geomorph. N.F., 30, 291-301p.

CROWELL, J.C. (1974): Origin of late Cenozoic basins in southern California. In W.R. Dickinson (ed.), Tectonics and Sedimentation. Tulsa, Special Publication Society Economic Paleontologists and Mineralogists, 22, 190-204p.

CURRIE, D.R. (1984): The derivation of 14C dating standards for fresh-water shell-fish. D.S.I.R. Institute of Nuclear Science, N.Z., INS-R 331.

DAVIS, W.M. (1902): "Base-level, grade, and peneplain". Journal of Geology, 10, 77-111p.

DONALD, R.C.; JOHN, D.V. (1978): Thresholds in Geomorphology. In Geomorphology Symposium. Binghampton, N.Y.

DOORNKAMP, J.C. (1986): Geomorphological approaches to the study of neotectonics. Journal of the Geological Society, 143, 335-342p.

DUPLESSEY, J.C. (1978): Climatic change. C.U.P.

EASTERBROOK, D.J. (1969): Principles of Geomorphology. McGraw-Hill.

EDGER, D.E. : Geomorphic and hydraulic properties of laboratory rivers. Colorado State University, 1973. (Thesis: M.Sc. 156p).

EMILIANI, C. (1972): Quaternary paleotemperatures and the duration of the high temperature intervals. Science, 178, 398-401p.

FAIR, E.E.: Structural, tectonic and climatic control of the fluvial geomorphology of the Manawatu River west of the Manawatu Gorge. Palmerston North, Massey University, 1968. (Thesis: M.Sc.: Geology).

FAIRBRIDGE, R.W. (1981): The concept of Neotectonic - An Introduction. Annals of Geomorphology, Suppl. Bd, 40.

FINDLAY, R.H. (1980): The Marshall Paraconformity. N.Z. Journal of Geology and Geophysics, 23, 125-134p.

FLEMING, C.A. (1975): The Quaternary record of New Zealand and Australia. In R.P. Suggate & M.M. Cresswell (eds), Quaternary Studies. Royal Society of New Zealand, Bulletin, 13, 1-20p.

FREUND, R. (1971): The Hope Fault: a strike-slip fault in New Zealand. Bulletin N.Z. Geological Survey, 86, 49p.

GAGE, M. (1970): Late Cretaceous and Tertiary rocks of Broken River, Canterbury. N.Z. Journal of Geology and Geophysics, 13, 507-559p.

GAGE, M. (1958): Late Pleistocene glaciation of the Waimakariri Valley, Canterbury, New Zealand. N.Z. Journal of Geology and Geophysics, 1, 135-155p.

GARDNER, T.W. (1975): The history of part of the Colorado River and its tributaries: An experimental study: Four corners. Geological Society Guidebook, 8th Field Conference (Canyonlands), 87-95p.

GARDNER, T.W.: A model study of river meander incision. Colorado State University, 1973. (Thesis: M.S.: Geology).

GHANI, M.A. (1978): Late Cenozoic vertical crustal movements in the southern North Island. N.Z. Journal of Geology and Geophysics, 21, 117-126p.

GIBB. J.G. (1986): A New Zealand regional Holocene eustatic sea-level curve and its application for determination of vertical tectonic movements. Royal Society of New Zealand, Bulletin, 24, 377-395p.

GIBB, J.G. : Late Quaternary shoreline movements in New Zealand. Wellington, Victoria University, 1979. (Thesis: Ph.D.: Geology).

GILEWSKA, S. (1967): Different methods of showing the relief on the detailed geomorphological maps. Zeitschrift fur Geomorphologie, n.f. 11, 481-490p.

GRAHAM, R.H. (1978): Wrench faults, arcuate fold patterns and deformation in the southern French Alps. Proceedings of the Geological Association, 89, 125-142p.

GREGG, D.R. (1959): Stratigraphy of the lower Waipara Gorge, North Canterbury. N.Z. Journal of Geology and Geophysics, 2, 501-527p.

GREGG, D.R.: Geology of Lower Waipara Gorge. Christchurch, University of Canterbury, 1950. (Thesis: M.Sc.: Geology).

GREGG, D.R. (1964): Sheet 18 Hurunui. Geological map of New Zealand 1:250,000. Wellington, N.Z.D.S.I.R.

HAAST, J. von (1872): Report of the geology of the Malvern Hills, Canterbury. N.Z. Geological Survey reports of geological explorations during 1871-1872, 7.

HAAST, J. von (1871b): On the geology of the Waipara District, Canterbury, with geological map and sections. N.Z. Geological Survey reports of geological explorations during 1870-1871, 6.

HACK, J.T. (1973a): Stream-profile analysis and stream gradient index. Journal of Research of U.S. Geological Survey, 1, 421-429p.

HACK, J.T. (1973b): Drainage adjustment in the Appalachians. In M.E. Morisawa (ed.), Fluvial Geomorphology. N.Y., Suny-Binghampton, 51-69p.

HACK, J.T. (1960): Interpretation of erosional topography in humid temperate regions. American Journal of Science, 258-A. Bradley Volume, 80-97p.

HACK, J.T. (1957): Studies of longitudinal stream profiles in Virginia and Maryland. U.S. Geological Survey, Professional Paper, 294-B, 45-97p.

HANKS, T.C.; WALLACE, R.E. (1985): Morphological analysis of the Lake Lahontan shoreline and beach front fault scarps, Pershing Country, Nevada. Bulletin Seismological Society of America, 75, 835-846p.

HANKS, T.C. and others. (1984): Modification of wave-cut and faulting-controlled landforms. Journal Geophysical Research, 89, 5771-5790p.

HARDING, T.P. (1974): Petroleum traps associated with wrench faults. American Association Petroleum Geologists, Bulletin, 58, 1290-1304p.

HARRIS, M.G.: Canterbury gravels, Omihi-stream section, Waipara. Dunedin, University of Otago, 1982. (Thesis: Dip. Sci.: Geology).

HARVEY, A.M.; WELLS, S.G. (1987): Response of Quaternary fluvial systems to differential epeirogneic uplift: Aguas and Feos river systmes, Southeast Spain. Geology, 15, 689-693p.

HAUSWIRTH, E.K.; SCHEIDEGGER, A.E. (1980): Tektonische Vorzeichung von Hangbewegungen im Raum von Bad Gastein, Interpravent Bd., 1, 159-178p.

HAUSWIRTH, E.K. and others. (1979): Untersuchungeneines Talzuschubes bei lesach (kals, Osttirol). Verh. Geol. B.A., 2, 51-79p.

HECTOR, J. (1884): Progress Report, 1883. N.Z. Geological Survey report of geological explorations during 1883-4, 16.

HOSKINS, R.H.: Planktonic foraminiferida and the Paleocene-Eocene boundary of the middle Waipara Gorge, Canterbury, New Zealand. Christchurch, University of Canterbury, 1969. (Thesis M.Sc.: Geology).

HULL, A.G.: Late Quaternary geology of the Cape Kidnappers area, Hawkes Bay, New Zealand. Wellington, Victoria University, 1985. (Thesis: M.Sc.: Geology).

HULL, A.G.; BERRYMAN, K.R. (1986): Holocene tectonism in the region of the Alpine Fault at Lake McKerrow, Fiordland, New Zealand. Royal Society of New Zealand, Bulletin, 24.

HULL, A.G. (in press): A late Holocene uplift shore platform on the Kidnappers coast, North Island, New Zealand: Some implications for shore platform development processes and uplift mechanism. Quaternary Research.

HUTCHINSON, J.N. (1968): "Mass Movement". In F.W. Fairbridge (ed.), Encyclopedia of Geomorphology. Reinhold, 688-695p.

HUTTON, F.W. (1905): The formation of the Canterbury Plains. Transactions of the New Zealand Institute, 37, 464-472p.

HUTTON, F.W. (1888): One some railway cuttings in the Weka Pass. Transactions of the Bew Zealand Institute, 20.

HUTTON, F.W. (1877): Report on the geology of the north-east portion of the South Island, from Cook Strait to the Rakaia. N.Z. Geological Survey report of geological explorations during 1872-3, 8b, 27-58p.

JENNINGS, J.N. (1971): Karst. Canberra, ANU Press.

JOBBERNS, G. (1937a): The Lower Waipara Gorge. Transactions of the Royal Society of New Zealand, 67, 125-132p.

JOBBERNS, G. (1937b): Alternative geomorphic histories of the North Canterbury landscape. Australia New Zealand Association for the Advancement of Science, Report, 23, 398p.

JOBBERNS, G. (1928): The raised beaches of the north east coast of the South Island of New Zealand. Transactions of the New Zealand Institute, 59, 508-570p.

JOBBERNS, G. (1926): Raised beaches in Teviotdale district, North Canterbury. Transactions of the New Zealand Institute, 56, 225-226p.

JOBBERNS, G.; KING, L.C. (1933): The nature and mode of origin of the Motunau Plain, North Canterbury, New Zealand. Transactions of the New Zealand Institute, 63, 355-369p.

KATZ, H.R. (1982): Petroleum prospects in North Cnaterbury, South Island, New Zealand. N.Z. Journal of Geology and Geophysics, 25, 201-207p.

KATZ, H.R. (1974): Margins of the Southwest Pacific. In C.A. Burke & C.L. Drake (eds), The Geology of Continental Margins. N.Y., Springer, 549-565p.

KELLER, E.A. and others. (1982a): Tectonic geomorphology of the San Andreas fault zone in the southern Indio Hills, Coachella Valley, California. Geological Society of America, Bulletin, 93, 46-56p.

KELLER, E.A.and otherrs. (1982b): Tectonic geomorphology of the Ventura, Ojai, and Santa Paula areas, Western Transverse Ranges, California. In Neotectonics of southern California. Geological Society of America, Cordilleran Section Meeting Guidebook, 25-42p.

KING, G; STEIN, R. (1983): Surface folding, river terrace deformation rate and earthquake repeat time in a reverse faulting environment. The 1983 Coalinga, California earthquakes. California Division of Mines and Geology, Special Publication, 66, 165-176p.

KOHN, B.P. (1979): Identification and significance of a late Pleistocene tephra in Canterbury District, South Island, New Zealand. Quaternary Research, 11, 78-92p.

LAUDER, W.R. (1962): Teschenites from Acheron River, mid-Canterbury, New Zealand, with notes on the geology of the surrounding country. Transactions of the Royal Society of New Zealand (Geology), 1, 109-127p.

LENSEN, G.J. (1968): Analysis of progressive fault displacement during downcutting at the Branch River Terraces, South Island. Geological Society of America, Bulletin, 79, 545-556p.

LEOPOLD, L.B.; LANGBEIN, W.B. (1962): The concept of entropy in landscape evolution. U.S. Geological Survey, Professional Paper, 500-A, 20p.

LEOPOLD, L.B.; MILLER, J.P. (1954): A post-glacial chronology for some alluvial valleys in Wyoming. U.S. Geological Survey Water Supply, Paper, 1261-1290p.

LEOPOLD, L.B.; MADDOCK, T. (1953): The hydraulic geometry of stream channels and some physiographic implications. U.S. Geolqocal Survey, Professional Paper, 252p.

LEWIS, D.W. (1976): Subaqueous debris flows of early Pleistocene age at Motunau, North Canterbury, New Zealand. N.Z. Journal of Geology and Geophysics, 19, 251-276p.

LEWIS, D.W. (1970): The New Zealand Zoophycos. N.Z. Journal of Geology and Geophysics, 13, 295-315p.

LEWIS, D.W.; BELLISS, S.E. (1984): Mid-Tertiary unconformities in the Waitaki subdivison, North Otago. Journal of the Royal Society of New Zealand, 14, 251-276p.

LEWIS, K.B. (1980): Quaternary sedimentation of the Hukurangi oblique-subduction and transform margin, New Zealand. International Association of Sedimentologists, Special Publication, 4, 171-184p.

LEWIS, W.V. (1944): Stream trough experiments and terrace formation. Geological Magazine, 81, 241-253p.

MacPHERSON, E.O. (1947): Stratigraphy of Chalk Hill, Oxford survey district, with notes on grading the chalk. N.Z. Journal of Science and Technology, 28(B), 164-172p.

MARDEN, M.: Late Pleistocene geology of the Kowai River valley, mid-Canterbury. Christchurch, University of Canterbury, 1976. (Thesis: M.Sc.: Geology).

MASON, B.H. (1949): The geology of the Mandamus-Pahau district, North Canterbury. Transactions of the Royal Society of New Zealand, 77, 403-428p.

MASON, B.H. (1941): The geology of Mount Grey district, North Canterbury. N.Z. Journal of Science and Technology, 29(B), 116-120p.

MAXWELL, P.A.: Structural geology and pre-Quaternary stratigraphy of the Kaiwara District, North Canterbury, New Zealand. Christchurch, University of Canterbury, 1964. (Thesis; M.Sc.: Geology).

MAYER, L. (1984): Dating Quuaternary fault scarps formed in alluvium using morphologic parameters. Quaternary Research, 22, 300-313p.

MAYER, L.: The geology and geomorphology of the Buckhead Mesa Area, Gila County, Arizona. Arizona, University of Arizona, 1979. (Thesis: M.Sc.: Geology).

MEANS, W.D. (1976): Stress and strain. N.Y., Springer-Verlag, 338p.

MILLER, R.L.; KAHN, J.S. (1962): Statistical analysis in the geological sciences. N.Y., Wiley, 470p.

MORISAWA, M.; HACK, J.T. (1985): Tectonic geomorphology. The Binghampton Symposia in Geomorphology: International Series, no.15, 390p.

MORRIS, L.J. (1983): Investigation of the Montserrat Landslide. Unpublished Report.

NASH, D.B. (1984): Morphologic dating of fluvial terrace scarps and fault scarps near West Yellowstone, Montana. Geological Society of America, Bulletin, 95, 1413-1424p.

NGS (1973): Reports on geodetic measurements of crustal movement, 1906-71. National Geodetic Survey, Washington, U.S. Department of Commerce.

OBERLANDER, T.M. (1985): Origin of drainge transverse to structures in orogens. In Morisawa, M. and Hack, J.T. (Eds.) "Tectonic Geomorphology", 155-182p.

OLIVER, P.J. (1984): The mid-Cretaceous volcanic rocks of Rakaia Gorge and Malvern Hills area, Canterbury. N.Z. Geological Survery, Record 3, 86-91p.

OTA, Y. (ed.) (1987): Holocene coastal tectonics of eastern North Island, New Zealand, 104p.

OTA, Y. and others. (1987): Height and age distributions of Holocene marine terraces on the south Wairarapa Coast, New Zealand. In Ota, Y. (ed), Holocene Coastal Tectonics of Eastern North Island, New Zealand, 60-78p.

OTA, Y. and others. (1981): Parts sheets O27, R27 and R28. Late Quaternary tectonic map of New Zealand 1:50,000. Wellington, N.Z. Geological Survey.

OTA, Y. and others. (1984): Marine terraces of the Conway Coast, South Island, New Zealand. N.Z. Journal of Geology and Geophysics, 27, 313-325p.

OUCHI, S. (1985): Response of alluvial rivers to slow active tectonic movement. Geological Society of America Bulletin. 96,504-515p.

PELTO, C.R. (1954): Mapping of multicomponent systems. Journal of Geology, 62, 501-511p.

PEPLIES, R.W.; KEUPER, H.F. (1975): Regional analysis. In R.G. Reeves (ed.), Manual of Remote Sensing. Virginia, American Society of Photogrammetry.

PILLANS, B. (1983): Upper Quaternary marine terrace chronology and deformation, South Taranaki, New Zealand. Geology, 11, 292-297p.

PONCE de LEON, M.I.; CHOUKROUNE, P. (1980): Shear zones in the Iberian Arc. Journal of Structural Geology, 2, 63-68p.

POWERS, W.E. (1962): Terraces of the Hurunui River, New Zealand. N.Z. Journal of Geology and Geophysics, 5, 114-129p.

REILLY, W.I. (1987): Crustal bending in New Zealand from first-order surveys 1925-1982. Discussion notes: Rapid Tectonics and Continental Collesion Workshop. New Zealand Geological Societ Annual Conference, Dunedin, December 1987.

ROBERTSON, A.H.F.; WOODCOCK, N.H. (1980): Strike-slip related sedimentation in the Antalya Complex, S.W. Turkey.
In P.F. Ballance & H.G. Reading (eds), Sedimentation in oblique-slip mobile zone, 127-145p.

RODGERS, D.A. (1980): Analysis of pull-apart basin development produced by en echelon strike-slip faults. In P.F. Ballance & H.G. Reading (eds), Sedimentation in oblique-slip mobile zone, 27-41p.

ROGERS, P.A.: Sedimentology and paleoecology of the upper Greta Siltstone, Motunau Beach, North Canterbury. Christchurch, University of Canterbury, 1970. (Thesis: B.Sc.Hons: Geography).

RUSS, D.P. (1982): Style and significance of surface deformation in the vicinity of New Madrid, Missouri. U.S. Geological Survey, Professional Paper, 1236-H, 95-114p.

RYNN, J.M.W.; SCHOLZ, C.H. (1978): Seismotectonics of the Arthurs Pass region, South Island, New Zealand. Geological Society of America, Bulletin, 89, 1373-1388p.

SALOME, A.I.; van DORSSER, H.J. (1985): Some reflections on geomorphological mapping systems. Z. Geomorph., 29, 375-380p.

SCHOLZ, C.H. and others. (1973): Detailed seismicity of the Alpine fault zone and Fiordland region, New Zealand. Geological Society of America, Bulletin, 84, 3297-3316p.

SCHUMM, S.A. (1981): Evolution and response of the fluvial system, sedimentologic implications. Society of Economic Paleontologists and Mineralogists, Special Publication, 31, 19-29p.

SCHUMM, S.A. (1979): Geomorphic thresholds: the concept and its applications. Institute of British Geographers, Transactions, 4, 485-515p.

SCHUMM, S.A. (1977): The Fluvial System. N.Y., John Wiley & Sons, 338p.

SCHUMM, S.A. and others. (1982): Phase 1: Investigation of neotectonic activity within the lower Mississippi Valley Division. Vicksburg, Missouri, 158p.

SCHUMM, S.A.; KHAN, H.R. (1972): Experimental study of channel patterns. Geological Society of America, Bulletin, 83, 1755-1770p.

SCOTT, R.M., and others. (1985): Land units of Chimbu province, Papua New Guinea. C.S.I.R.O., Natural Resources Series, 5, 162p.

SEEBER, L.; GORNITZ, V. (1983): River profiles along the Himalayan arc as indications of active tectonics. Tectonophysics, 92, 335-367p.

SELBY, M.J. (1985): Earth's changing surface. An introduction to geomorphology. Oxford, Clarendon Press, 607p.

SEWELL, R.J.: The volcanic geology and geochemistry of central Banks Peninsula and relationships to Lyttelton and Akaroa volcanoes. Christchurch, University of Canterbury, 1985. (Thesis: Ph.D.: Geology).

SCHOLZ, C.H. and others. (1973): Detailed seismicity of the Alpine Fault Zone and Fiordland region, New Zealand. Geological Society of America, Bulletin, 84, 3297-3316p.

SHACKLETON, N.J.; MATTHEWS, R.K. (1977): Oxygen isotope stratigraphy of Late Pleistocene coral terraces in Barbadoes. Nature, 268, 618-620p.

SHACKLETON, N.J.; OPDYKE, N.D. (1973): Oxygen-isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: oxygen isotope temperatures and ice volumes on a 10^5 and 10_6 year scale. Quaternary Research, 3, 39-55p.

SHANNON, C.E.; WEAVER, W. (1949): The mathematical theory of communication. Urbana, University of Illinois Press.

SHEPHERD, R.G. (1985): Regression analysis of river profiles. Journal of Geology, 93, 377-384p.

SMALE, D.; and others (1982): A bedding-plane land-slide near Mt Vulcan, North Canterbury. N.Z. Journal of Geology and Geophysics, 25, 397-404p.

SMART, G.M.: The geology of the Mount Cass District, North Canterbury, New Zealand. Christchurch, University of Canterbury, 1954. (Thesis: M.Sc.: Geology).

SMITH, W.D. (1983): Seismology and related research in New Zealand 1979-1982. Seismological Observatory, Bulletin, 276, 32p.

SNOW, R.S.; SLINGERLAND, R.L. (1987): Mathematical modelling of graded river profiles. Journal of Geology, 95, 15-33p.

SOONS, J.M. (1963): The glacial sequence in part of the Rakaia Valley, Canterbury, New Zealand. N.Z. Journal of Geology and Geophysics, 6, 735-756p.

SOONS, J.M.; SELBY, M.J. (1982): Landforms of New Zealand. Longman Paul Ltd, 392p.

SOONS, J.M.; BURROWS, C.J. (1978): Dates for Otiran deposits including plant microfossils and macrofossils, from Rakaia Valley. N.Z. Journal of Geology and Geophysics, 21, 607-615p.

SOONS, J.M.; GULLENTOPS, F.W. (1973): Glacial advances in the Rakaia Valley, New Zealand. N.Z. Journal of Geology and Geophysics, 16, 425-438p.

SPEIGHT, R. (1933): The Rakaia Valley. Transactions of the New Zealand Institute, 63, 457-496p.

SPEIGHT, R. (1919): The older gravels of North Canterbury. Transactions of the New Zealand Institute, 51, 269-281p.

SPEIGHT, R. (1915): The intermontane basins of Canterbury. Transactions of the New Zealand Institute, 47, 336-353p.

SPEIGHT, R. (1912): A preliminary account of the lower Waipara Gorge. Transactions of the New Zealand Institute, 44, 221-223p.

SPEIGHT, R.; WILD, L.J. (1918): The stratigraphical relationship of the Weka Pass Stone and the Amuri Limestone. Transactions of the New Zealand Institute, 50, 65-93p.

STEARNS, C.E. (1976): Estimates of the position of sea level between 140 and 75 thousand years. Quaternary Research, 6, 445-449p.

STEARNS, R.G. (1967): Warping of the western highland rim peneplain in Tennessee by groundwater sapping. Geological Society of America, Bulletin, 78, 1111-1124p.

STERR, H. (1985): Rates of change and degradation of hillslopes formed in unconsolidated materials: A morphometric approach to date Quaternary scarps, north-central Nevada. Geological Society of America, Bulletin, 88, 1267-1281p.

STRONG, C.P. (1984): Cretaceous-Tertiary boundary, mid-Waipara River section, North Canterbury, New Zealand (Note). N.Z. Journal of Geology and Geophysics, 27, 231-234p.

SUGGATE, R.P. (1978): The Kaikoura Orogeny. In R.P. Suggate and others (eds), The Geology of New Zealand. Wellington, Government Printer, N.Z., 671-701p.

SUGGATE, R.P. (1974): When did the last interglacial end? Quaternary Research, 4, 246-252p.

SUGGATE, R.P. (1965): Late Pleistocene geology of the northern part of the South Island, New Zealand. Geological Survey of New Zealand, Bulletin, 77, 91p.

SWEETING, M.M. (1972): Karst Landforms. London, Macmillan.

TATOR, B.A. (1958): The aerial photograph and applied geomorphology. Photogrametric Engineering, 24, 549-561p.

THOMSON, J.A. (1920): The Notocene geology of the middle Waipara and Weka Pass District, north Canterbury, New Zealand. Transactions of the New Zealand Institute, 52, 322-415p.

THORNBURY, W.D. (1969): Principles of geomorphology. 2nd ed. John Wiley & Sons, 594p.

THORNES, J.B.; BRUNSDEN (1977): Geomorphology and Time. Great Britain, Halsted Press, 208p.

Van DORSSER, H.J.; SALOME, A.I. (1973): Different methods of detailed geomorphological mapping (with coloured example). K.N.A.G. Geogr. Tijdschr. N.S.7., 71-74p.

VARNES, D.J. (1978): Slope movements, types and processes. In R.L. Schuster & R.J. Krizek (eds), Landslides, analysis and control. Special Report 176, Washington D.C. National Academy of Science.

VERSTAPPEN, H. Th. (1977): Remote sensing in geomorphology. Amsterdam, Elsevier. 214p.

VERSTAPPEN, H. Th.; van ZUIDAM, R.A. (1975): ITC system of geomorphological survey. 3rd. ed. In ITC textbook of photo-interpretation, VII.2. The Netherlands, ITC. 52p.

VERSTAPPEN, H.Th. (1970): Introduction to the ITC-system of geomorphological survey. Geogr. Tijdschr. 4(1), 85-91p.

VERSTAPPEN, H.Th; van ZUIDAM, R.A. (1968): ITC system of geomorphological survey. In ITC textbook of photo-interpretation, VII.2. The Netherlands, ITC. 52p.

VOLKOV, N.G. and others. (1967): Effect of recent crustal movements on the shape of longitudinal profiles and water levels in rivers. In Symposium on River Mechanics, Bern: International Union Geodesy and Geophysics Publication, 75, 105-116p.

WALCOTT, R.I. (1984): The kinematics of the plate boundary zone through New Zealand: a comparison of short- and long-term deformation. Geophys. Jour. Res. astr. Soc. 79, 613-633p.

WALCOTT, R.I.; BIBBY, H.M. (1984): The kinematics of a subduction and transform plate boundary: The Pacific-Australian plate boundary zone through New Zealand. Bulletin of the Royal Society of New Zealand.

WALCOTT, R.I. and others. (1980): Bending within the axial tectonic belt of New Zealand in the last 9 MYR from paleomagnetic data. Earth & Planetary Science Letters 52, 427-434p.

WALLACE, R.E. (1978): Geometry and rates of change of fault-generated range fronts, north-central Nevada. U.S. Geological Survey, Journal of Research, 6, 637-650p.

WALLACE, R.E. (1977a): Profiles and ages of young fault scarps, north central Nevada. Geological Society of America, Bulletin 88, 1267-1281p.

WARDS, I. (ed) (1976): New Zealand Atlas. Wellington, Government Printing Office, 292p.

WARREN, G.; SPEDEN, I.G. (1978): The Piripauan and Haumurian stratotypes (Mata Series, Upper Cretaceous) and correlative sequences in the Haumuri Bluff district, South Marlborough (S56). N.Z. Geological Survey Bulletin, 92.

WATSON, C.C. : An assessment of the Lower Mississippi River below Natchez, Mississippi. Unpublished Ph.D. thesis (1982), Colorado State University, 162p.

WAY, D.S. (1973): Terrain analysis; a guide to site selection using aerial photographic interpretation. Stroudsburg, Pennsylvania, Dowden Hutchinson & Ross, 392p.

WEBB, P.N. (1966): New Zealand Late Cretaceous foraminifera and stratigraphy. Utrecht, Schotanus & Jens, 19p.

WELCH, D.M. (1973): Channel form and bank erosion, Red River, Manitoba. In Hydrological Symposium, 9th Edmonton, Proceedings. Alberta, Canada, University of Alberta, 284-293p.

WILCOX, R.E. and others. (1973): Basic wrench tectonics. American Association of Petroleum Geologists, Bulletin, 57, 74-96p.

WILSON, D.D. (1963): Geology of the Waipara Subdivision. N.Z. Geological Survey Bulletin, 64, 122p.

WILSON, D.D. (1955): The northern continuation of the Canterbury Plains, Canterbury, New Zealand. N.Z. Journal of Science and Technology, 37, 126-131p.

WOODFORD, A.D. (1951): Stream gradients and Monterey sea valley. Geological Society of America, Bulletin, 62, 799-852p.

YEATS, R.S. and others (1981): Active fault hazard in southern California: Ground rupture versus seismic shaking. Geological Society of America, Bulletin, 92, 189-196p.

YOUSIF, H.S. (in press): Geomorphological approaches to the study of neotectonics: lower Waipara gorge, North Canterbury, New Zealand. N.Z. Journal of Geology and Geophysics.

YOUSIF, H.S. (1986): The evolution of alluvial systems in response to active folding in North Canterbury. (Abstract). Australia & New Zealand Geomorphology Group, Napier Conference, Feb. 1986.

YOUSIF, H.S. (1984): Tectonic geomorphology of the lower Waipara gorge. (Abstract). Geological Society of New Zealand, Victoria University of Wellington Conference, Dec. 1984, Miscellaneous Publication, 31A.

ZUIDAM, R.A. van; ZUIDAM, C.F.I. van (1979): Terrain analysis and classification using aerial photographs. In ITC textbook of photo-interpretation. The Netherlands, ITC. 310p.

APPENDIX 1

MORPHOMETRIC ANALYSIS

1.1 INTRODUCTION

A major part of this appendix has involved the preparation of 63 longitudinal stream profiles that have served as the basis for various interpretative maps and diagrams. Computer data used in the preparation of stream profiles has come from the recent work by the writer using the digitzer device.

1.2 PROGRAM TO CALCULATE LONGITUDINAL STREAM PROFILES

Symbols used in the longitudinal profiles and tables:

K : The gradient index of the entire profile.

GI = SL : The gradient index of a specific reach.

GG : The gradient index of a middle reach.

LOG (GI): The logarithm of gradient index of a specific reach.

L : The distance (m).

X & Y : The coordinates of elevation points.

```
10 CLS:SCREEN 0,0:WIDTH 80
30 KEY OFF
50 REM   **********************************************
70 REM
90 REM   *    PROGRAM TO CALCULATE LONGITUDINAL          *
110 REM   *            STREAM PROFILES                    *
130 REM   *                                               *
150 REM   **********************************************
170 REM
190 REM
200 DEFDBL R,G
210 PRINT "**************************************************"
230 PRINT " "
250 PRINT " This program calculates LONGITUDINAL STREAM PROFILES"
270 PRINT " "
290 PRINT "**************************************************"
310 PRINT " "
312 INPUT "Name of Output file: ",DISKFIL$
315 OPEN DISKFIL$ FOR OUTPUT AS #3
320 INPUT "CATCHMENT AREA: ",CATCH$
325 LPRINT USING "CATCHMENT AREA: & ";CATCH$:LPRINT:PRINT #3,"CATCHMENT AREA ",CATCH$;
330 INPUT " RIVER NO. ",RN:INPUT "GRID REFERENCE ",GX
350 LPRINT USING "& #########    & #########";" RIVER NO. ",RN,"  GRID REFERENCE ",GX:LPRINT
360 PRINT #3,"RIVER NO :",RN,"GRD REFERENCE ",GX
370 INPUT "FIRST ELEVATION: ",HI:INPUT "LAST ELEVATION :",HJ
380 PRINT #3,"FIRST ELEVATION ",HI, "LAST ELEVATION ",HJ,
390 LPRINT USING "&  ######";" FIRST  ELEVATION (FT) ",HI,"  LAST ELEVATION (FT) ",HJ,"  DIFFERENCE IN ELEVATION ",HI-HJ
410 LPRINT:LPRINT
430 LPRINT "POINT       X          Y          DELTA L       LENGTH(L)        L(m.)    GRAD INDEX FOR SEGMENT    LOG(GI)       GG"
450 PRINT "THE X AND Y COORDINATES ARE ENTERED BY PRESSING THE EN"
470 PRINT "BUTTON ON THE CURSOR."
490 PRINT "TO START PRESS ANY KEY ON COMPUTER KEYBOARD"
510 G$ =INKEY$
530 IF G$="" THEN 510
550 DIM X(50),Y(50),DL(50),L(50),G(50),GI(50),GG(50)
570 J=0
590 MENU =5 'Value of the MENU-KEY is ctrl-E
610 CLS
630 CLOSE#1
650 OPEN "COM2:1200,e,7,2,rs,cs0,ds0,cd0,lf,asc" AS #1
670 OPEN "scrn:" FOR OUTPUT AS #2
690 GOSUB 1730
710 COM(2) OFF
```

```
730 CLS:PRINT #2,"INPUT DATA"
750 COM(2) ON
770 ON COM(2) GOSUB 1250
790 B$=INKEY$
810 IF B$<>"" THEN 850
830 GOTO 790
850 IF ASC(B$)=MENU THEN 910
870 PRINT #1,B$;:PRINT #2,B$;
890 GOTO 790
910 LPRINT:LPRINT
930 REM TO CALCULATE THE GRADIENT INDEX OF THE ENTIRE PROFILE
950 K=(HI-HJ)/(LOG(L(J))-(LOG(L(2))))
970 LPRINT USING "THE GRADIENT INDEX OF THE ENTIRE PROFILE ,K = ######.####";K
980 PRINT #3, "THE GRADIENT INDEX OF THE ENTIRE PROFILE,K= ",K,
990 LPRINT:LPRINT
1010 LPRINT " POINT      GRAD.INDEX/K "
1030 FOR I=2 TO J-1
1050 LPRINT USING "####         ####.#######";I,GI(I)/K
1060 PRINT #3, "POINT ",J, "GRAD INDEX /K ",GI (I)/K;
1070 NEXT I
1090 LPRINT CHR$(12)
1095 REM
1100 GOSUB 2000:REM PRINT A PLOT OF THE PROFILE
1105 REM
1110 CLOSE #1,#2,#3
1130 KEY ON
1150 END
1170 REM
1190 REM  ******* PROCESS DATA FROM DIGITABLET *******
1210 REM
1230 REM
1250 LINE INPUT #1,A$:C$=A$
1270 IF (MID$(C$,6,1)<> ".") OR (MID$(C$,10,1)<>",") OR (MID$(C$,16,1)<>".") THEN 1610
1290 J=J+1
1310 PRINT #2 ,"POINT ",J
1330 PRINT #2," X and Y values are :"
1350 X$=MID$(C$,2,8):Y$=MID$(C$,13,8)
1370 PRINT #2,X$,Y$
1390 X(J)=VAL(X$):Y(J)=VAL(Y$)
1410 IF J<2 THEN LPRINT USING "####   ####.###   ####.###";J,X(J),Y(J)
1420 PRINT #3,  J,X(J),Y(J)
1430 IF J <2 THEN  1670
1450 DL(J)=SQR((X(J)-X(J-1))^2+(Y(J)-Y(J-1))^2)
1470 L(J)=L(J-1)+DL(J)
1490 IF J<3 THEN LPRINT USING "####   ####.###   ####.###   ##.#######   ###.#######   #######.##";J,X(J),Y(J),DL(J),L(J),L(J)*250
1500 PRINT #3,J,X(J),Y(J),DL(J),L(J),L(J)*250
```

```
1510 IF J<3 THEN 1590
1530 GI(J-1)=100/(LOG(L(J))-LOG(L(J-1)))
1550 GG(J-1)=100*.5*(L(J-1)+L(J))/(L(J)-L(J-1))
1570 LPRINT USING "####   ####.###   ####.###   ##.#######   ###.#######   #######.##   ####.####   #########   ##.#######   ####.#### "
;J,X(J),Y(J),DL(J),L(J),L(J)*250,GI(J-1),J-1,LOG(GI(J-1)),GG(J-1)
1580 PRINT #3,J,X(J),Y(J),DL(J),L(J),L(J)*250,GI(J-1),J-1 LOG(GI(J-1)),GG(J-1)
1590 GOTO 1670
1610 G$=LEFT$(C$,1):IF G$<"0" OR G$>"9" THEN GOTO 1670
1630 INPUT "CHANGE ",O$
1640 PRINT #3, "CHANGE ",O$;
1650 LPRINT:LPRINT O$:LPRINT
1670 A$="":RETURN
1690 REM ### TO SET UP DIGITABLET ###
1710 REM
1730 COM(2) ON
1750 ON COM(2) GOSUB 1950
1790 PRINT #2,STRING$(80,32)
1810 PRINT #2,"Set ORIGIN ":PRINT #1,"ORIGIN"+CHR$(13);
1830 PRINT #2,"Strike ANY key when FINISHED"
1850 G$=INKEY$
1870 IF G$="" THEN 1850
1880 PRINT #1,CHR$(67)+CHR$(77)+CHR$(13)+CHR$(10);
1885 FOR I=1 TO 1200:NEXT I
1890 RETURN
1910 REM **** OUTPUT FROM DIGITABLET ***
1930 REM
1950 LINE INPUT #1,C$:PRINT #2,C$:C$=""
1970 RETURN
1990 REM SUBROUTINE TO PRINT A PLOT OF PROFILE
2000 DIM TC%(50):TC%(0)=5
2010 DOT$="*":TX=12
2012 FOR K=1 TO J
2015 L(K)=L(K)*250
2017 NEXT K
2020 LPRINT :LPRINT
2030 LPRINT TAB(40),"LONGITUDINAL  STREAM PROFILES"
2040 LPRINT :LPRINT
2050 LPRINT "CATCHMENT AREA:  ";CATCH$;"        RIVER NO:  ";RN
2055 LPRINT "GRID REFERENCE :  ";GX
```

```
2060 LPRINT :LPRINT
2070 UNIT =(L(J)-L(1))/100
2075 MM=HI
2080 FOR K=2 TO J
2090 I=J-K+2
2100 LPRINT TAB(TX);"     I"
2110 LPRINT TAB(TX);"     I"
2120 LPRINT TAB(TX-2);MM;:IF MM> 999 THEN LPRINT "I";: GOTO 2140
2130 LPRINT " I";
2140 PCC =(L(K)-L(K-1))/UNIT
2150 PCC%=INT(PCC)
2160 TC%(K)=TC%(K-1)+PCC%
2170 LPRINT SPC(TC%(K))DOT$
2180 MM=MM-100
2190 NEXT K
2200 LPRINT TAB(TX);"     I":LPRINT TAB(TX);"     I"
2210 LPRINT TAB(TX+4);
2220 FOR N=1 TO 115
2230 LPRINT "_";
2240 NEXT N
2250 LPRINT:LPRINT TAB(TX);
2260 FOR N=2 TO J
2270 KK%=TC%(N)-TC%(N-1)-2
2280 LPRINT;SPC(KK%);L(N);
2290 NEXT N
2295 LPRINT CHR$(12)
2300 RETURN
```

1.3 ANALYSED DATA

Microfiche copies of the tables and longitudinal stream profiles are given in Appendix 1 at the back pocket of this thesis.

APPENDIX 2

SURVEYING ANALYSIS

2.1 INTRODUCTION

This appendix represents data necessary for the presentation of a river centreline profile of the Waipara River (from State Highway 1 to the river mouth), and terrace and strath profiles associated with the work undertaken by the author. The survey was carried out during May, June and July, 1984, by the author and A. McCare.

The survey of the Carrington Creek together with the strath and terrace surfaces was carried out during September, 1985.

2.2 FIELD PROCEDURE

Two procedures were used corresponding to the river centreline profile and river terrace profiles.

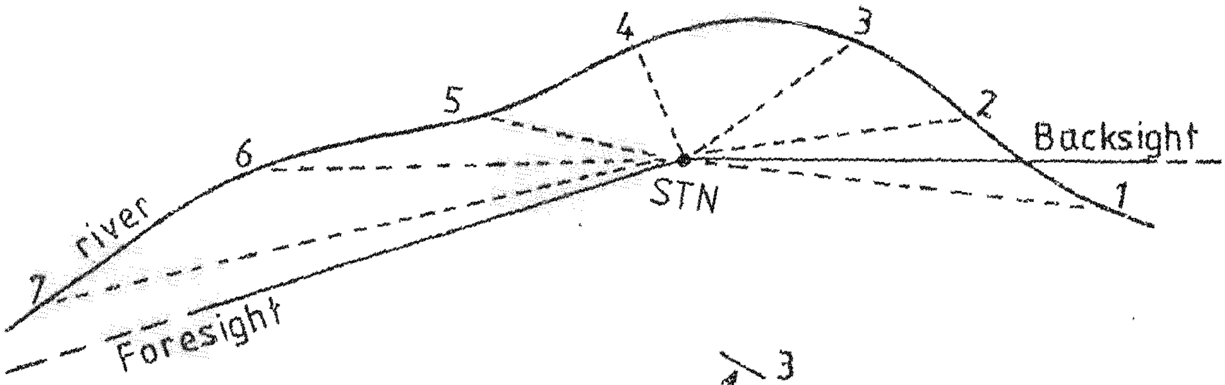
A. River Centreline Profile

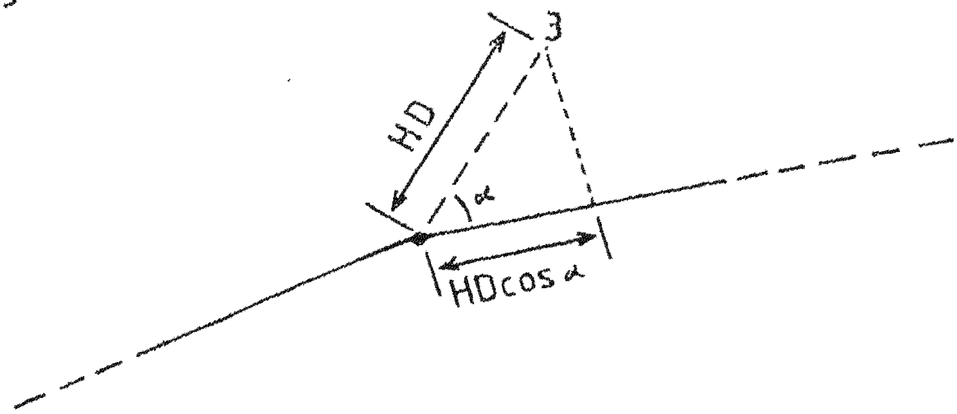
(1) Set up instrument at particular station concerned, recording height of instrument above station.

(2) Set up reflector at previous station, recording height of reflector above station.

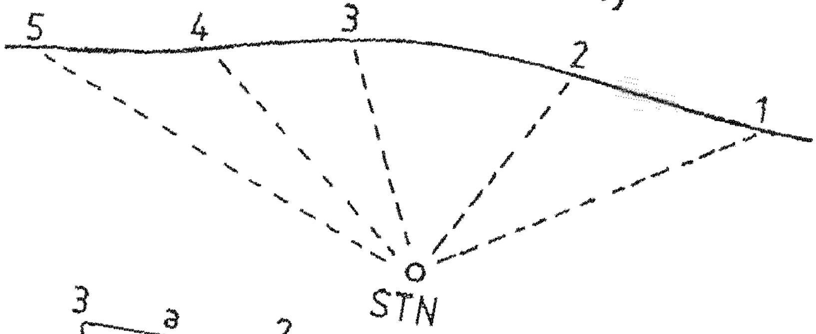
(3) Backsight to previous station, setting horizontal bearing and renording vertical angle, distance, and elevation difference.

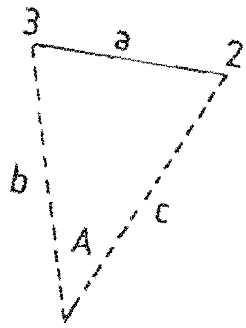
River centreline survey





Strath, terrace survey





$$a = \sqrt{b^2 + c^2 - 2bc\cos A}$$

(4) Sight to intermediate spot levels for river centreline profile, recording

- horizontal bearing
- vertical angle
- distance
- elevation difference

(5) Set up reflector at next station, recording reflector height.

(6) Foresight to next station recording

- horizontal bearing
- vertical angle
- distance
- elevation distance

Setting horizontal bearing: This entails taking the horizontal bearing from the previous station to the station at which the instrument is set up, less 180^{0} e.g. Foresight from Station RCS3 - RCS4 = $274^{0}40'00"$

set up at RCS4.

Backsight setting horizontal bearing

RCS4 - RCS3 = $94^{0}40'00"$

Sample Field Sheet (RCS4)

B. <u>River Terrace and Strath Profiles</u>

(1) Set up instrument at a river centreline station, or in a position to sight a river centreline station and terrace or strath to be observed.

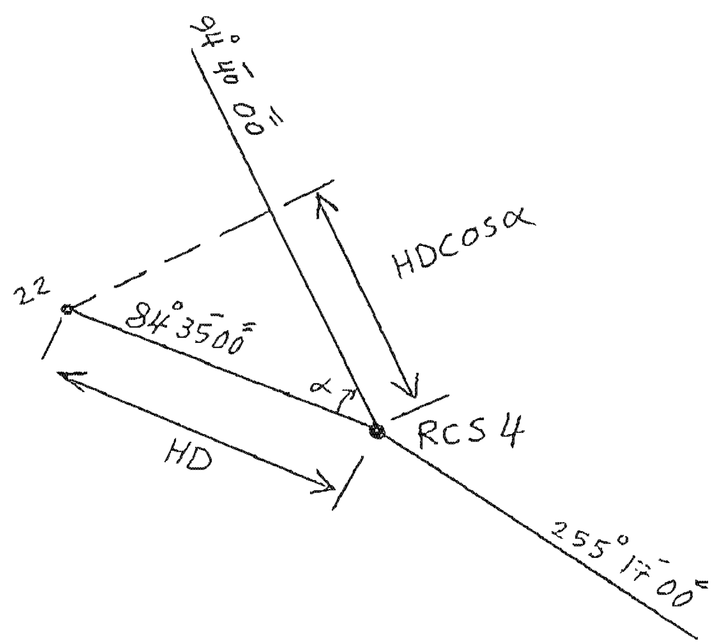
(2) Determine instrument height by either

(i) Measure height above river centreline station, or

SURVEY WAIPARA RIVER
RIVER BED PROFILE
LEVELS & DISTANCE DOWN RIVER

STATION RCS4	R.L. OF STATION	52.57	HEIGHT OF INSTRUMEN	1.63	R.L. OF INSTRUMENT	54.20	INSTTRUME T: Wild DI3. T1A

VERTICAL CIRCLE READING	SLOPE DISTANCE	STAFF HEIGHT	HEIGHT DIFFERENCE	REDUCED LEVEL	BEARING	HORIZONTAL DISTANCE (m)	POINT OR STATION SIGHTED	a	ACCUMULATED DISTANCE (m)		SYMBOL KEY
270 11 30	264.379	1.620	0.885	53.46	94 40 00	264.377	RCS3				B/S
269 36 50	76.167	1.69	-0.513	52.00	100 45 00	76.166	21	26.62	653.18	R	US
268 49 30	49.846	1.69	-1.022	51.49	84 35 00	49.836	22	16.5	679.80		
268 26 00	40.014	1.69	-1.093	51.42	59 11 00	40.000	23	32.27	696.30		
266 58 00	21.904	1.69	-1.171	51.34	347 18 00	21.873	24	38.17	728.57		
268 05 00	43.639	1.69	-1.459	51.05	283 01 30	43.615	25	38.16	766.74		
268 30 00	79.228	1.69	-2.073	50.44	271 33 36	79.201	26	39.59	804.90		
269 00 00	116.293	1.69	-2.029	50.48	261 22 30	116.275	27	24.84	844.49	R	DS
269 35 00	193.022	1.60	-1.403	51.20	255 17 00	193.017	RCS5				F/S



Point	α	HD	Cosα HD	DIST	RL
21	6.08	76.166	75.74	653.18	52.00
22	10.08	49.834	49.07	679.80	51.49
23	35.48	40.00	32.57	696.30	51.42
24	90	00	00	728.57	51.34
25	29.74	43.615	37.87	766.74	51.05
26	16.28	79.201	76.03	804.90	50.44
27	6.09	116.275	115.62	844.49	50.48

(ii) backsight to river centreline station for level only, recording

- vertical angle
- distance
- elevation difference

(3) Sight to terrace points recording

- horizontal bearing
- vertical angle
- distance
- elevation difference

2.3 CALCULATIONS

A. River Centreline Profile

River centreline stations form a river centreline traverse, and the distance downstream of a particular point is related to this traverse in the following manner.

Refering to the sample fieldsheet and calculation sheet (RCS4)

The component of the line running from the station RCS4 to the point No.22, in the traverse direction is determined thus:

HD = 49.836m (From Field Sheet)

= 10.08^{0} ($94^{0}40'00'' - 84^{0}35'00''$)

HDCOS α = 49.836 x COS 10.08'

HDCOS α = 49.07 m.

This distance is then subtracted from the distance for station RCS4, thus giving the distance for the point 22.

Distance of point 22 = 728.87 - 49.07 = 679.80 m.

The procedure beyond the station is the same except noting that the component of the line running from station to point in the traverse direction is added.

For points such as point 24, where the angle from the traverse is greater than 90^0, the distance of the station is assigned to the point.

All calculations for the river centreline profile were performed by hand using Hewlitt-Packard 11C.

B. <u>River Terrace and Strath Profiles</u>

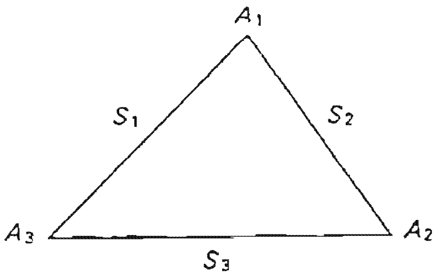
Calculations entailed determining the distance between consecutive points along the terrace or strath profiles. This was done using a Hewlitt-Packard HP 11C program based on the triangle equation

$a^2 = b^2 + c^2 - 2bcCosA$

Paramteres b,c,A are recorded in the field.

C. Triangle Solutions

This program may be used to find the sides, the angles, and the area of a plane triangle.



In general, the specification of any three of the six parameters of a triangle (3 sides, 3 angles) is sufficient to define a triangle. (The exception is that three angles will not define a triangle). There are thus five possible cases that this program will handle: two sides and the included angle (SAS), two angles and the included side (ASA), two sides and the adjacent angle (SSA—an ambiguous case), two angles and the adjacent side (AAS), and three sides (SSS).

If the three known input values are selected in a clockwise order around the triangle, the outputs will also follow a clockwise order.

Remarks:

- Inputs may be in any angular mode (i.e., DEG, RAD, GRAD). Be sure to set the angular mode of the calculator to match that of the inputs.
- Note that the triangle described by the program does not conform to standard triangle notation; i.e., A_1 is not opposite S_1.
- Angles must be entered as decimals. The [g] [→H] conversion can be used to convert degrees, minutes, and seconds to decimal degrees.
- Accuracy of solution may degenerate for triangles containing extremely small angles.

KEYSTROKES	DISPLAY	KEYSTROKES	DISPLAY
STO 3	058– 44 3	f LBL D	087–42,21,14
R↓	059– 33	STO 3	088– 44 3
STO 1	060– 44 1	R↓	089– 33
RCL 3	061– 45 3	STO 2	090– 44 2
RCL 4	062– 45 4	R↓	091– 33
SIN	063– 23	STO 1	092– 44 1
RCL 1	064– 45 1	f LBL 0	093–42,21, 0
÷	065– 10	RCL 2	094– 45 2
×	066– 20	RCL 1	095– 45 1
g SIN⁻¹	067– 43 23	f →R	096– 42 26
RCL 4	068– 45 4	RCL 3	097– 45 3
+	069– 40	x≷y	098– 34
GSB 9	070– 32 9	−	099– 30
STO 2	071– 44 2	g →P	100– 43 26
GSB 0	072– 32 0	STO 5	101– 44 5
RCL 1	073– 45 1	x≷y	102– 34
RCL 3	074– 45 3	STO 4	103– 44 4
f x≤y	075– 42 10	RCL 2	104– 45 2
GTO 8	076– 22 8	+	105– 40
2	077– 2	GSB 9	106– 32 9
R/S	078– 31	STO 6	107– 44 6
RCL 6	079– 45 6	SIN	108– 23
GSB 9	080– 32 9	×	109– 20
STO 6	081– 44 6	RCL 1	110– 45 1
RCL 4	082– 45 4	×	111– 20
+	083– 40	2	112– 2
GSB 9	084– 32 9	÷	113– 10
STO 2	085– 44 2	STO 0	114– 44 0
GTO 0	086– 22 0	.	115– 48

KEYSTROKES	DISPLAY	KEYSTROKES	DISPLAY
[f] CLEAR [PRGM]	000-	[RCL] 2	029- 45 2
[f] [LBL] [A]	001-42,21,11	[RCL] 6	030- 45 6
[STO] 5	002- 44 5	[+]	031- 40
[R↓]	003- 33	[SIN]	032- 23
[STO] 3	004- 44 3	[÷]	033- 10
[R↓]	005- 33	[RCL] 1	034- 45 1
[STO] 1	006- 44 1	[×]	035- 20
[RCL] 3	007- 45 3	[STO] 3	036- 44 3
[g] [→P]	008- 43 26	[GTO] 0	037- 22 0
[g] [x^2]	009- 43 11	[f] [LBL] [C]	038-42,21,13
[RCL] 5	010- 45 5	[STO] 4	039- 44 4
[g] [x^2]	011- 43 11	[R↓]	040- 33
[−]	012- 30	[STO] 2	041- 44 2
[RCL] 1	013- 45 1	[R↓]	042- 33
[RCL] 3	014- 45 3	[STO] 1	043- 44 1
[×]	015- 20	[RCL] 4	044- 45 4
2	016- 2	[RCL] 2	045- 45 2
[×]	017- 20	[+]	046- 40
[÷]	018- 10	[SIN]	047- 23
[g] [COS^{-1}]	019- 43 24	[RCL] 4	048- 45 4
[STO] 2	020- 44 2	[SIN]	049- 23
[GTO] 0	021- 22 0	[÷]	050- 10
[f] [LBL] [B]	022-42,21,12	[RCL] 1	051- 45 1
[STO] 2	023- 44 2	[×]	052- 20
[R↓]	024- 33	[STO] 3	053- 44 3
[STO] 1	025- 44 1	[GTO] 0	054- 22 0
[R↓]	026- 33	[f] [LBL] [E]	055-42,21,15
[STO] 6	027- 44 6	[STO] 4	056- 44 4
[SIN]	028- 23	[R↓]	057- 33

KEYSTROKES	DISPLAY	KEYSTROKES	DISPLAY
0	116- 0	[GTO] 2	124- 22 2
0	117- 0	[f] [LBL] 8	125-42,21, 8
6	118- 6	[g] [CLx]	126- 43 35
[STO] [I]	119- 44 25	[g] [RTN]	127- 43 32
[f] [LBL] 2	120-42,21, 2	[f] [LBL] 9	128-42,21, 9
[RCL] [(i)]	121- 45 24	[COS]	129- 24
[R/S]	122- 31	[CHS]	130- 16
[f] [ISG]	123- 42 6	[g] [COS^{-1}]	131- 43 24

REGISTERS			R_I: Index
R_0: Area	R_1: S_1	R_2: A_1	R_3: S_2
R_4: A_2	R_5: S_3	R_6: A_3	R_7—R_8: Unused

STEP	INSTRUCTIONS	INPUT DATA/UNITS	KEYSTROKES	OUTPUT DATA/UNITS
1	Key in the program.			
2	Set User mode.			
3	Set desired angular mode			
	(DEG, RAD, GRAD).			

STEP	INSTRUCTIONS	INPUT DATA/UNITS	KEYSTROKES	OUTPUT DATA/UNITS
4	SSS [3 sides]			
	Enter:			
	Side 1	S_1	[ENTER]	
	Side 2	S_2	[ENTER]	
	Side 3	S_3	[A]	Area
	Go to step 9.			
5	ASA [2 angles + included side]			
	Enter:			
	Angle 3	A_3	[ENTER]	
	Side 1	S_1	[ENTER]	
	Angle 1	A_1	[B]	Area
	Go to step 9.			
6	SAA [2 angles + adjacent side]			
	Enter:			
	Side 1	S_1	[ENTER]	
	Angle 1	A_1	[ENTER]	
	Angle 2	A_2	[C]	Area
	Go to step 9.			
7	SAS [2 sides + included angle]			
	Enter:			
	Side 1	S_1	[ENTER]	
	Angle 1	A_1	[ENTER]	
	Side 2	S_2	[D]	Area
	Go to step 9.			
8	SSA [2 sides + adjacent angle]			
	Enter:			
	Side 1	S_1	[ENTER]	
	Side 2	S_2	[ENTER]	

STEP	INSTRUCTIONS	INPUT DATA/UNITS	KEYSTROKES	OUTPUT DATA/UNITS
	Angle 2	A_2	[E]	Area
9	Dimensions:			
	To find the remaining sides and			
	angles:		[R/S]	S_1
			[R/S]	A_1
			[R/S]	S_2
			[R/S]	A_2
			[R/S]	S_3
			[R/S]	A_3
9a	If no alternative triangle is possible:		[R/S]	0
9b	If second triangle is possible:		[R/S]	2
			[R/S]	Area
			[R/S]	S_1
			[R/S]	A_1
			[R/S]	S_2
			[R/S]	A_2
			[R/S]	S_3
			[R/S]	A_3

2.4 SURVEY EQUIPMENT

1. Wild D13 Distomat with Wild T1A Theodolite
2. Special DI3 Reflector, the GDR 31.

2.5 ANALYSED DATA

Microfich copies of the survey data of the river centreline profiles, strath and terrace profiles, of both the lower Waipara River and the Carrington Creek are given in Appendix 2 at the back pocket of this thesis.

APPENDIX 3

RADIOCARBON DATING

Microfiche copies of the radiocarbon racord forms and results are given in Appendix 3 at the back pocket of this thesis.

APPENDIX 4

LANDSAT IMAGES

Microfiche copies of the Landsat imagery utilized in this research in the form of black and white, enlarged, enhanced and positive prints are given in Appendix 4 at the back pocket of this thesis.

APPENDIX 5

PUBLICATIONS

1. YOUSIF, H.S. 1984: Tectonic geomorphology of the lower Waipara gorge (Abstract).

2. YOUSIF, H.S. 1986: The evolution of alluvial systems in response to active folding in North Canterbury (Abstract).

3. CAMPBELL, J.K.; YOUSIF, H.S. (1985): Tectonic geomorphology of the lower Waipara gorge, North Canterbury (published field guides).

4. YOUSIF, H.S. (in press): Geomorphological approaches to the study of neotectonics: Lower Waipara Gorge, North Canterbury, New Zealand Journal of Geology and Geophysics (see also Chapter 6 section 6.3).

Microfiche copies of these publications are given in appendix 5 at the back pocket of this thesis.